

Long-term Global Morphology of Gravity Wave Activity Using UARS Data

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ABSTRACT

This report provides a broad outline of the total body of research conducted during the past three years. We report on detailed model studies of the precise way in which infrared limb scanning satellites explicitly detect gravity waves, and use these insights to resolve earlier discrepant zonal mean estimates of gravity wave variances from satellite limb scanners. Detailed analysis of CRISTA temperature fluctuations are outlined, which provide new global information on long-wavelength stratospheric gravity waves generated by mountains, tropical convection and the mid-latitude jet stream-vortex system, as well as interactions higher in the mesosphere with the diurnal tide. The detailed insights gained from analyzing CRISTA data are applied to provide a multi-year analysis of stratospheric mountain waves over the Andes evident in MLS limb-track data. We also demonstrate for the first time that stratospheric temperature data from the CLAES instrument on UARS resolved gravity waves, and we highlight mountain wave information in a subset of these data. Final conclusions and recommendations are set forth, and the present findings are directly related to the original goals of the research contract. A full list of publications that resulted from this research is provided.

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1. Introduction

Following the guidelines outlined in the proposal contract, we present a final report on the project without explicitly reporting on results performed during the last three months of the project just completed. We have performed a wide variety of tasks under this contract, and it is impossible to outline all the results in this report: much of it has been documented in the previous collection of eleven quarterly reports submitted during the three-year project lifetime.

Newer work that has yet to be written up will receive the major writeups in this report. Work that has been successfully transitioned to be publication stage will be reviewed more briefly.

This research has benefited greatly from collaborations forged during the three years of this project. I particularly wish to acknowledge the pivotal role played by Dr. Peter Preusse, of the Department of Physics and Mathematical Physics, University of Wuppertal, Germany, whose dedicated efforts made much of the work outlined here possible.

2. Major Work and Results (Including New Results This Quarter)

2.1 Are Gravity Waves Resolved by Infrared Limb-Scanning Satellites?

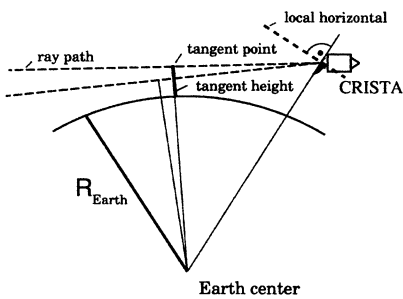


Figure 1: Measuring geometry of the limb sounder. Two rays with different tangent heights are shown. CRISTA orbits clockwise in this Figure, whereas for CLAES or MLS the orbit is either perpendicularly into or out of the page (depending on the yaw cycle).

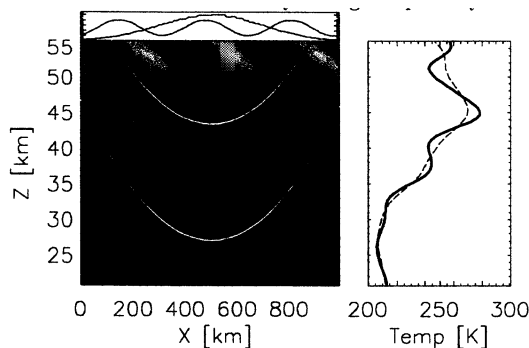


Figure 2: Viewing geometry of a limb sounding projected through a simulated two-dimensional gravity wave temperature perturbation superimposed upon a background temperature profile. Two limb ray paths with different tangent heights are shown in left panel. Top panel gives the weighting function along a ray path as well as wave oscillation phase

Our work in modeling the way in which CRISTA and other limb scanning instruments detect gravity waves has been set out at length in previous reports [Preusse *et al.*, 1999; section 1.4 of report dated March 18-June 18, 1998; section 1.4 of June 18-September 18, 1998; section 1.2 of report dated 18 June-September 18, 2000]. Since the bulk of the work has been reported previously, only a brief overview of the major results and some previously unreported results will be provided here. The major findings are that CRISTA explicitly resolves long wavelength gravity waves from retrieved temperatures derived from inversion of $12.6 \mu\text{m}$ CO_2 limb-scanned radiances. These results also apply to similar temperature data acquired by the CLAES instrument on UARS. The gravity waves extracted from both instruments' data are discussed in sections 2.3 and 2.5.

The observational situation is shown schematically in Figure 1. In our model, we insert a linear two-dimensional sinusoidal temperature fluctuation due to a gravity wave, projecting those oscillations into a 2D plane both in altitude, z , and in horizontal distance along the limb-scan direction, x . This oscillation exists on top of a mean background temperature profile. Then we simulate a limb scan of the atmosphere by CRISTA under the influence of this perturbed temperature profile, yielding $12.6 \mu\text{m}$ CO_2 radiance profiles, of the form

$$I(z) = \int_{-\infty}^{\infty} \xi(s) B[T(s)] ds$$

The geometry of this situation is set out in Figure 2, showing limb ray paths through a two-dimensional gravity wave. Here s is the distance increment along one of the limb ray paths shown in Figure 2, $\xi(s)$ is

the emissivity at distance s along the limb ray path, and $B(T)$ is the blackbody radiance function.

From here, we passed these model limb scanned radiance data $I(z)$ through the same retrieval algorithms used for actual CRISTA retrievals, to back out a temperature profile $T(z)$. Then we analyzed the fluctuations that remain on the extracted profile and compared the measured amplitude of the fluctuation, \hat{T} , to the actual amplitude of the wave we inserted, \tilde{T} . This yields a “visibility,” $\varepsilon = \hat{T}/\tilde{T}$, of CRISTA to the gravity wave of this particular amplitude with the given vertical and horizontal

wavelength pairing (λ_h, λ_z) . We then conducted a series of these simulations using different gravity wave wavelengths λ_h and λ_z to infer a net two-dimensional CRISTA visibility function, $\varepsilon(\lambda_h, \lambda_z)$ to gravity waves with varying wavelength properties aligned along the limb-scan direction.

These calculations are extremely expensive to do numerically, and a considerable amount of computing time and effort was required to generate results that are summarized in Figures 3b and 3c. The major advance in these new calculations over previous more limited efforts is the use of full BANDPAK libraries [Marshall *et al.*, 1994] to generate realistic radiance fields measured along the limb, and use of updated CRISTA retrieval algorithms in the inversion back to a limb-scanned temperature profile. Figure 3b shows results in which the amplitudes of the temperature perturbations in Figure 2 were kept constant at 1 K. Figure 3c shows similar collated results in which the temperature amplitudes were allowed to grow quasi-exponentially with altitude. Figure 3a shows an analytical simplification in which we simplified the problem by assuming that:

- The blackbody radiance function $B(T)$ can be approximated by the first term in its Taylor series expansion
- The considered range of wavelengths is optically thin.
- The atmosphere is isothermal
- The wave amplitude remains constant with altitude

Further details of the calculations and the analytical simplification are provided by Preusse *et al.* [2000c].

In summary, we see that the three sets of plots in Figure 3 are broadly similar. All show >50% visibility of CRISTA to gravity waves with vertical wavelengths $\lambda_z > 5$ km and $\lambda_h > 200$ km. Below these limits, CRISTA’s visibility characteristics diminish

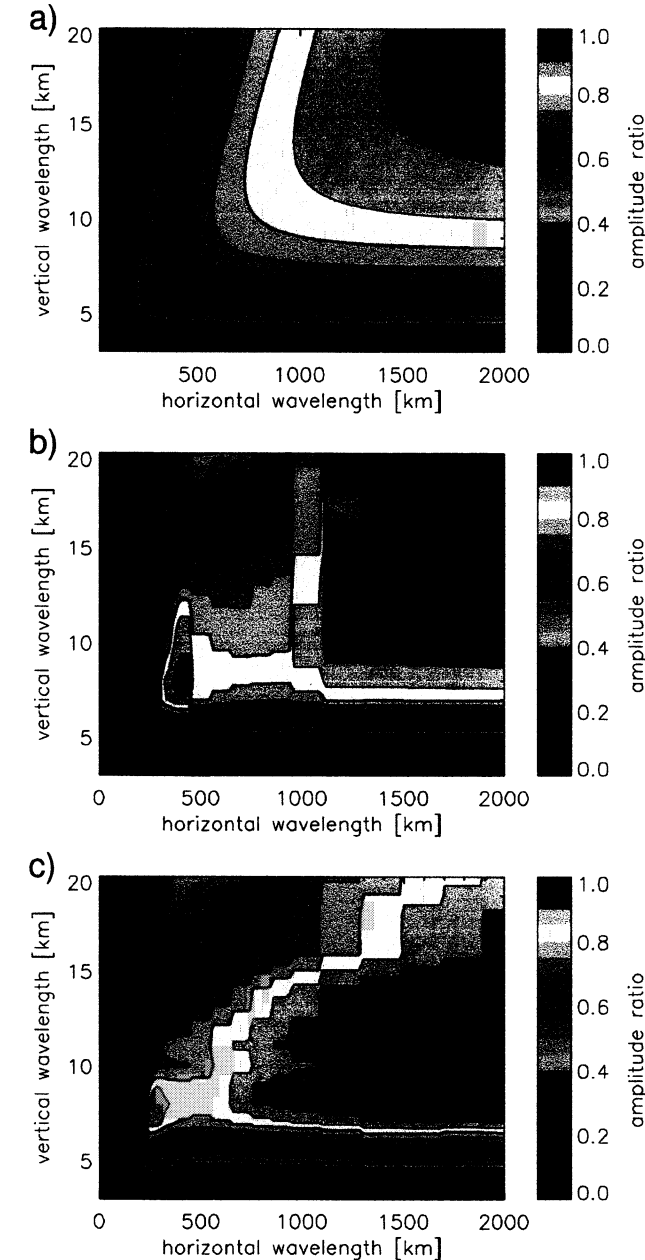


Figure 3: Estimated visibilities $\varepsilon(\lambda_h, \lambda_z)$ of CRISTA temperatures to gravity waves with different horizontal and vertical wavelengths. Radiative transport and well as retrieval effects are included in the calculations. Panel (b) shows full calculations using a constant amplitude of 1K, (c) using amplitudes that grow exponentially with height. Panel (a) shows an analytical model for the visibility function that is developed under certain simplifications by Preusse *et al.* [2000c].

rapidly, indicating strong smearing of the wave response due to integration of radiance fluctuations

along the limb. We view these two-dimensional simulations as “worst case” scenarios, since here CRISTA scans the limb at a 90° angle to the wave phase fronts, as shown in Figure 4 (top). If CRISTA scanned at some other angle, as shown schematically in Figure 4 (bottom), then the smearing of the temperature wave along the limb is much reduced, since the wave presents a longer horizontal wavelength component along the limb direction. Thus, we believe that the zones of $>50\%$ visibility to gravity waves in Figure 3 mark those long wavelength gravity waves that CRISTA should be able to observe nearly all the time, regardless of the horizontal orientation of the phase fronts with respect to the CRISTA viewing direction. Waves with lower visibility will tend not to be observed in these “worst case” scenarios, but may be episodically visible if/when CRISTA scans the wave in some configuration closer to the “best case” scenario in Figure 4. Indeed, this can be gauged directly from Figure 3. As depicted schematically in Figure 5, for an angle α between the CRISTA scan direction and the gravity wave horizontal wavevector direction, the component of the horizontal wavenumber intercepted by the CRISTA ray path scanned through the line of sight (LOS) along the limb (pink curve in Figure 4) is $k_{\text{limb}} = k_h \cos \alpha$, where $k_h = 2\pi/\lambda_h$ is the gravity wave’s horizontal wavenumber and α is 0° or 180° for the “worst case” and 90° or 270° for the “best case.” Thus the wavelength along the limb LOS resolved by CRISTA is $\lambda_{\text{limb}} = \lambda_h \sec \alpha$, so that for the “best case” scenario in Figure 4, λ_{limb} is essentially infinite and the visibility of CRISTA to the gravity wave in this case is given by values for a given λ_z at the long horizontal wavelength limit of Figure 3.

In addition to the fidelity of the amplitude retrieval, we have also studied how the wave’s vertical wavelength and phase are affected by the retrieval. Figure 6 shows a scatterplot of the actual versus the retrieved vertical wavelengths. The effects here combined both the weaknesses in the retrieval as well as the MEM/HA analysis to isolate gravity waves in temperature profiles (this method that has been discussed at length in previous reports and will be discussed briefly in section 2.3.1). We see from Figure 6 generally good retrieval of the vertical wavelength of the gravity wave in each case. We also investigate the absolute phase of the retrieved disturbance at a given altitude. The combined results are shown in Figure 7. We see that phase shifts are generally small, except at the boundary regions of the amplitude sensitivity evident in Figure 3b. Thus, as expected, the wave phase retrieval is less reliable in those regions of Figures 3b and 7 where the wave is at the limit of the instrument’s intrinsic ability to “see” the gravity wave. We have also studied the altitude dependence of these gravity wave retrievals. Figure 8 shows a plot of the one-

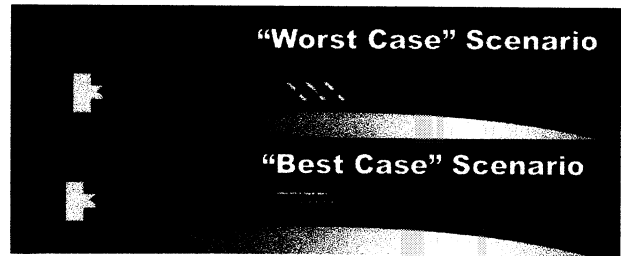


Figure 4: Idealized schematic showing “worst case” ($\alpha = 0^\circ$ or 180°) and “best case” scenarios ($\alpha = 90^\circ$ or 270°) for CRISTA limb scan through a gravity wave segment, shown as red-yellow temperature banding in a quasi—horizontal plane near the tangent height. Figure 2 represents the physics of the “worst case” scenario above.

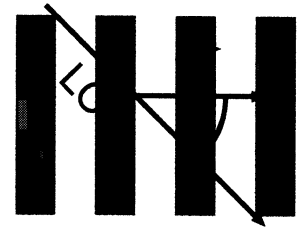


Figure 5: Relation between CRISTA line of sight (LOS), shown in pink in Figure 4, wave phase fronts, and the gravity wave horizontal wavenumber vector \vec{k}_h and the

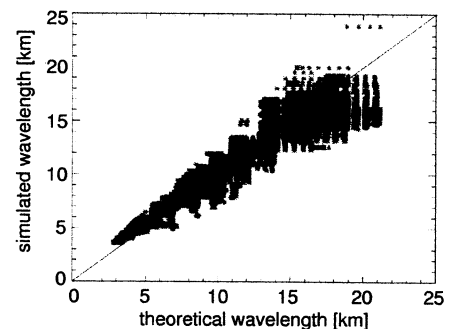


Figure 6: Scatterplot of retrieved versus input vertical wavelength. Solid line indicates 100% accurate retrieved value.

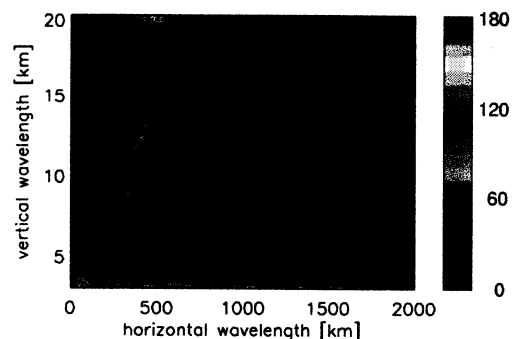


Figure 7: As for Figure 3b, but showing phase shifts in retrieved compared to actual wave.

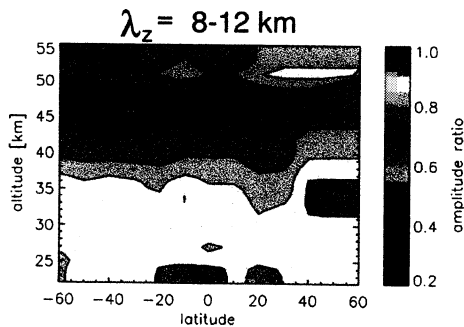


Figure 8: Latitude-height distribution of one-dimensional mean visibility function $\hat{\epsilon}(\lambda_z)$ in the 8-12 km vertical wavelength range.

dimensional visibility approximation (see section 2.2) inferred over the 8-12 km range of vertical wavelengths. We see there is little latitudinal dependence to the visibility, but some slight height dependence, with marginally better sensitivity at lower altitudes compared to higher stratospheric altitudes.

These results were formulated specifically for observations with the CRISTA instrument, since we have full details of the data, the radiance measurements, the retrieval algorithms and the data analysis algorithms. Nonetheless, we believe that these results also translate to temperature measurements acquired by the CLAES instrument on UARS, which acquires its stratospheric temperature data via inversion of limb scanned radiances from the $12.6 \mu\text{m}$ CO_2 Q-branch emissions [see, e.g., Gille *et al.*, 1996]. Thus, the results in Figures 3-8 immediately suggested to us that we should be able to resolve gravity waves in CLAES temperatures, in much the same way that we have verified both theoretically and experimentally for CRISTA. This work motivates the study in section 2.5.

Although we have provided some depth to the analysis here, we cannot provide full details. Initial results can be found in the publications of Preusse *et al.* [1999] and Preusse *et al.* [2000a] (the latter reproduced in section 7.2), and a full writeup of the modeling work outlined herein is provided in the work of Preusse *et al.* [2000c]. Another separate spectral study of a special high-resolution subset of CRISTA data has also largely supported the theoretical sensitivity results presented here. That work of ours has just been submitted for publication [Eidmann *et al.*, 2000] but is not described here due to space constraints.

2.2 Reconciling Zonal Mean Global Estimates of Gravity Wave Temperature Variances in the Middle Atmosphere Measured by Various Satellite Instruments

At the beginning of this project, some preliminary maps of small-scale temperature variance had been compiled from satellites. Examples include analysis of small-scale temperature fluctuations in data from the Limb Infrared Monitor of the Stratosphere (LIMS) on the NIMBUS 7 satellite [Fetzer and Gille, 1994], and radiance fluctuations from microwave limb radiance data from the Microwave Limb Sounder (MLS) instrument on the UARS satellite [Wu and Waters, 1997]. Zonal mean maps of variances from these two instruments are shown in Figure 9. It was noted that the activity differed greatly between these and other instruments. These discrepancies, along with the fact that gravity wave fluctuations existed at the spatial and precision limits of all these instruments, meant that there was little agreement as to the

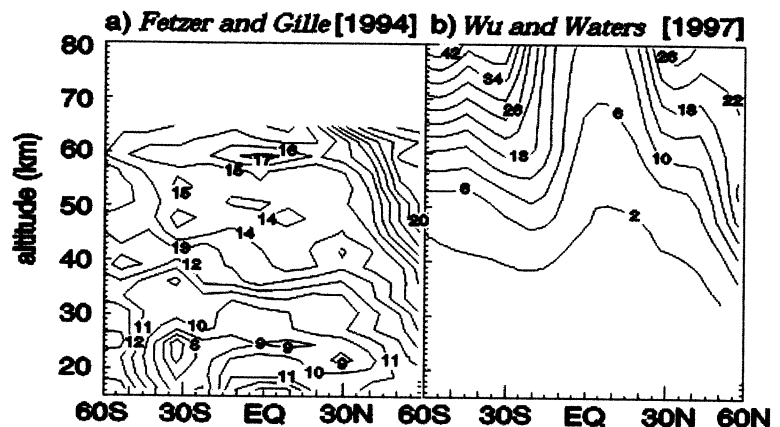


Figure 9: Zonal mean distributions of small-scale temperature-related variance due to gravity waves, plotted as a function of latitude and height, from satellite data acquired by (a) LIMS [after Fetzer and Gille, 1994] and (b) MLS [after Wu and Waters, 1997]. LIMS contours are temperature variances in dB over 1 K^2 (i.e., 10 dB = 10 K^2 , 20 dB = 100 K^2). MLS contours are relative radiance variances. Both datasets were acquired during northern winter (6-12 November, 1978 for LIMS data, December 20, 1992-January 29, 1993 for the MLS data).

geophysical significance of these preliminary satellite-derived gravity wave data. Indeed, at the beginning of this project, it was unclear whether any of these instruments were reliably measuring gravity waves. A major focus of this project was to investigate these issues both theoretically and observationally.

The modeling study outlined in section 2.1 revealed that limb-scanning satellites should, theoretically, be able to resolve long wavelength gravity waves in the stratosphere and mesosphere. Why, then, are zonal mean results like those in Figure 9 so variable among the various instruments?

Alexander [1998] used model ray tracing data to propose that much of the difference in Figure 9 could be explained by the spatial sensitivity functions of the various instruments. Put simply, the two-dimensional visibility functions (see Figure 3) can vary considerably between instruments, meaning that each instrument is observing different wavelength regions of the total gravity wave field in the middle atmosphere. She used these ideas to filter model data to resemble the wavelength scales resolved by MLS and various ground-based instruments such as rockets and radiosondes.

We conducted a study to test these ideas experimentally using CRISTA data as a reference, given the well-defined gravity wave visibility and radiance characteristics we have inferred for this instrument [Riese *et al.*, 1999; see also section 2.1]. First, we approximated the two-dimensional CRISTA visibilities $\varepsilon(\lambda_h, \lambda_z)$ in Figure 3 by a

one-dimensional visibility term, $\hat{\varepsilon}(\lambda_z)$, derived by averaging $\varepsilon(\lambda_h, \lambda_z)$ over the horizontal wavelength range from 300-800 km. This is largely valid since, from Figure 3, the visibility does not vary significantly over this range, but falls off rapidly below it. We can now use this one-dimensional visibility factor, $\hat{\varepsilon}(\lambda_z)$, to correct for imperfect observations by CRISTA of the measured gravity wave amplitudes \hat{T} without having to know what the wave's horizontal wavelength was (horizontal wavelengths are difficult to measure from satellite instruments given the comparatively coarser horizontal resolution of satellite data).

Next we applied new one-dimensional visibility functions to these corrected CRISTA data that crudely represented those of the MLS, LIMS and GPS/MET instruments, represented here by the functions $\hat{\varepsilon}_{MLS}(\lambda_z)$, $\hat{\varepsilon}_{LIMS}(\lambda_z)$, and $\hat{\varepsilon}_{GPS}(\lambda_z)$, respectively. In each case, following Alexander [1998], we adopted simple boxcar visibility functions with passbands in the range $\lambda_z > 10$ -25 km for MLS below 60 km, $\lambda_z > 15$ -35 km for MLS above 60 km (where the weighting function is much broader), $\lambda_z > 3$ -15 km for LIMS, and $\lambda_z > 3$ -

10 km for GPS/MET. These functions were applied to the corrected MEM/HA \hat{T} data from CRISTA, to derive new CRISTA datasets that were “MLS-like,” “LIMS-like,” and “GPS/MET-like.” For more background and details on these choices, see Preusse *et al.* [2000a].

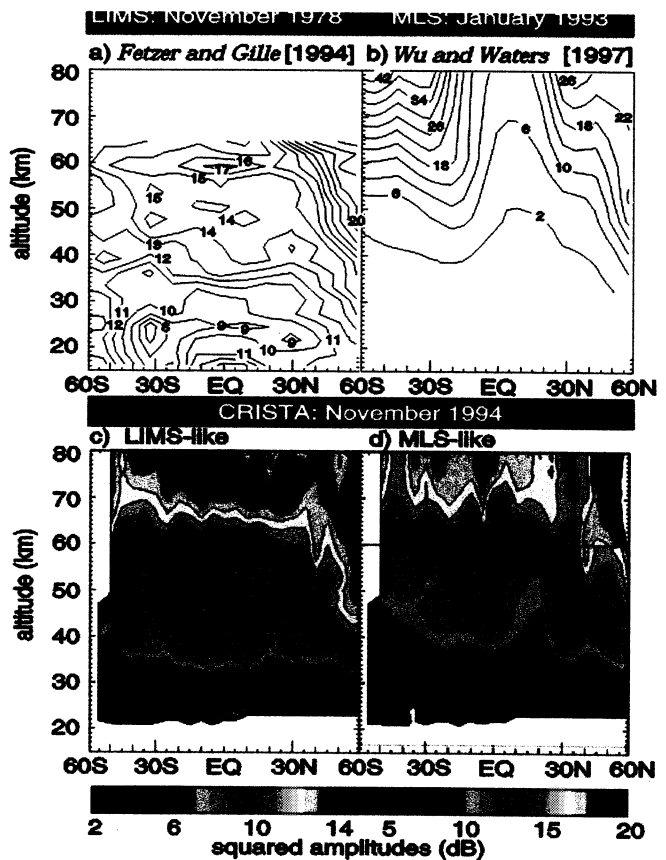


Figure 10: Top row as in Figure 9. Bottom row shows “LIMS-like” and “MLS-like” zonal-mean \hat{T} data from CRISTA. See text for further details. Line in (d) separates high-altitude and low-altitude estimates, which use different one-dimensional visibility functions.

Latitude-height global zonal-mean results for the resulting “MLS-like” and “LIMS-like” CRISTA data are shown in Figures 10c and 10d, beneath the original data reproduced from Figure 9 to facilitate easier visual comparisons. We see that the two filtered CRISTA datasets are quite different from each other, due to the effects of the boxcar $\hat{\epsilon}_{MLS}(\lambda_z)$ and $\hat{\epsilon}_{LIMS}(\lambda_z)$ visibility functions applied to the corrected data. Furthermore, we see features that resemble the actual LIMS and MLS data redisplayed above in Figures 10a and 10b.

For example, for the LIMS comparisons, we see at low levels that Figure 10c reproduces the equatorial maximum and the minima at 30°S and 30°N evident in the actual data in Figure 10a. Note that these features are not nearly as evident in the MLS-like CRISTA data shown in Figure 10d. We also see a “ledge” of reduced variance growth, or even reductions, with height over the 40-60 km altitude range. Indeed, similar things are also seen in gravity wave temperature data derived from rocketsondes, which have similar vertical wavelength sensitivity [Eckermann *et al.*, 1995].

For the MLS comparisons (right column Figure 10), we see in the “MLS-like” CRISTA data in Figure 10d the general reproduction of a band of attenuated variance centered at ~20°N, which is seen clearly in the actual MLS data reproduced in Figure 10b. The “MLS-like” CRISTA variance grows monotonically with altitude, showing little if any sign of the ledge at 40-60 km in the LIMS data and the “LIMS-like” CRISTA data. Again, the same monotonic increase with altitude is evident in the MLS data in Figure 10b.

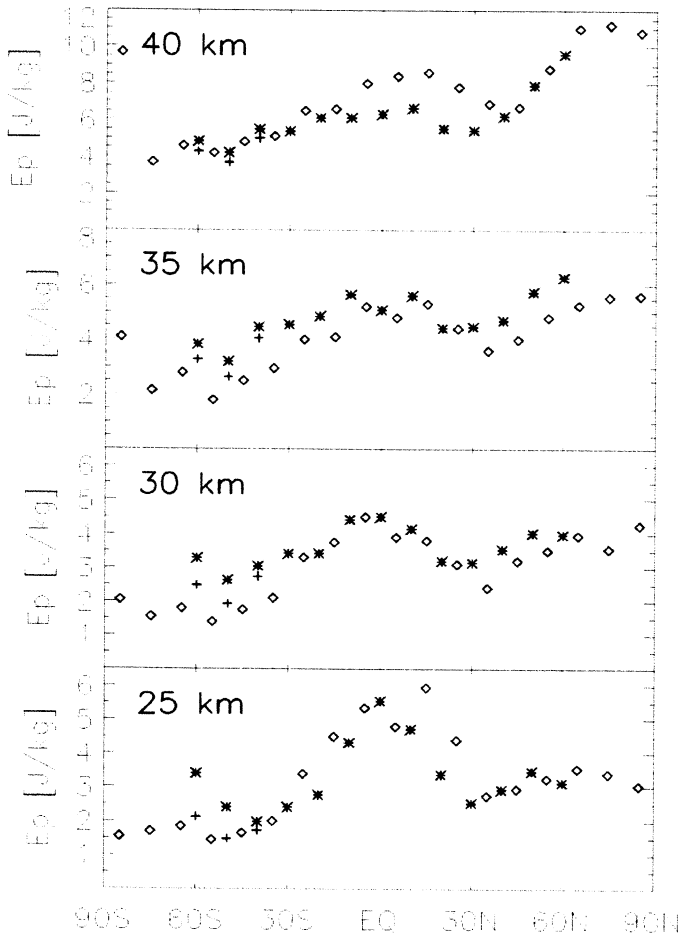


Figure 11: Zonal mean gravity-wave potential energies per unit mass, as a function of latitude and height, derived from GPS/MET data (asterisks) and “GPS/MET-like” CRISTA MEM/HA data (asterisks). Crosses near 40-60S show recalculated “GPS/MET-like” CRISTA MEM/HA data after removal of large values over the Andes.

We also used these ideas to compare with recent gravity wave data inferred from temperature profiles provided by GPS occultation measurements made from the GPS/MET satellite system [Tsuda *et al.*, 2000]. Latitude height variations inferred from the wintertime GPS/MET data are plotted with open diamonds in Figure 11, where here the temperature variances have been converted to potential energies per unit mass using standard gravity wave formulas [see Tsuda *et al.*, 2000]. Corresponding results from the “GPS/MET-like” CRISTA \hat{T} data are plotted with asterisks in Figure 11. We see excellent correspondence between the two, both in shape of latitudinal variations at each altitude, and in absolute magnitudes. At 25 km, the CRISTA data show a larger peak near 40°-60°S due to large mountain waves over the southern Andes in the CRISTA data (see sections 2.3.2 and Figures 14, 15, 17, 20, 21, and 28). When we remove data from the Andean sector of longitudes, we get the new results shown with crosses in Figure 11. Once again, the data here correspond quite well with the mean GPS/MET values of Tsuda *et al.* [2000].

The results in Figures 10-11 and elsewhere provide direct experimental verification of the weighting function argument originally postulated by Alexander

[1998] using model data. They show that temperature perturbation amplitudes \hat{T} from different instruments resemble each other once the wavelength filtering characteristics of the measurements are taken into account. It also shows that CRISTA, MLS, LIMS and GPS/MET all made reliable measurements of gravity waves within the particular portions of the gravity wave spectrum they were sensitive to. These results, as well as those in section 2.1, achieve one of the major goals of the project – to assess whether high-resolution satellite instruments on UARS and other satellites make reliable measurements of gravity wave oscillations. Figures 10 and 11 prove beyond doubt that they do.

This work has been described in greater detail in previous reports [e.g., section 1.1 of the report dated December 18, 1998-March 18, 1999; section 1.3 of report dated June 18-September 18, 2000]. A short version of this work was recently published in *Geophysical Research Letters* [Preusse et al., 2000a – see section 7.2] and more complete analyses and details are provided in a paper being drafted for the special issue of *The Journal of Geophysical Research* devoted to first results from the CRISTA 2 mission [Preusse et al., 2000c].

Given the fidelity of the gravity wave signals that sections 2.1 and 2.2 have set forth, the following sections describe detailed analysis of the gravity wave signals in data from different satellite instruments.

2.3 Gravity Wave Activity in CRISTA Temperatures Acquired During the CRISTA-SPAS Satellite Missions

The CRISTA instrument flew on CRISTA-SPAS (Shuttle Pallet Satellite) for two missions of ~1 week duration during shuttle missions STS-66 (4-12 November, 1994) and STS-85 (7-16 August, 1997). CRISTA data taken during each mission will hereafter be referred to as CRISTA 1 and CRISTA 2 data, respectively. Here we analyze CRISTA temperatures acquired in the stratosphere from 12.6 μm CO₂ emissions, and mesospheric data from the 15 μm emission. Analysis in previous sections shows that these data contain gravity wave-scale spatial structures. We infer amplitudes \hat{T} of small-scale temperature fluctuations due to gravity waves in these CRISTA data using the MEM/HA analysis method applied to Kalman-detrended global temperature fields for each day of the mission, as developed and outlined in previous reports. We outline the final procedure in a formal way in section 2.3.1. We use the \hat{T} data from this analysis of the 12.6 μm CRISTA temperature channel to study stratospheric gravity waves emanating from mountains (section 2.3.2), from the jet-vortex region (section 2.3.3) and from tropical convection (section 2.3.4). We conclude by analyzing high altitude temperature data from the 15 μm CRISTA temperature channel to study mesospheric gravity waves in section 2.3.5.

2.3.1 Extraction of Gravity Waves from Global CRISTA Temperature Data

Over the course of this contract, we developed and refined our own, largely new method for extracting gravity wave temperature fluctuations from global satellite data, which we describe only briefly here, since it has been described at length in various other places [Preusse et al., 1999; Eckermann and Preusse, 1999; Preusse et al., 2000c]. Nonetheless, since it is a key analysis method used throughout the project, we describe it briefly in this section.

First, given global asynchronously sampled temperature data each day, we remove the planetary-scale structure using a Kalman filter [Rodgers, 1977] that fitted temperatures at each level to zonal wavenumbers 0-6, which should account for most of the mean state and planetary waves that exist globally [e.g., Fetzer and Gille, 1994]. In addition, we fit data from ascending and descending nodes separately, since these data are separated in local time by ~12 hours and thus these separate fits help to remove the diurnal tide (this is important in the mesosphere – see section 2.3.5). The background temperature estimated by the Kalman fits at various altitudes are interpolated to the precise locations of each temperature measurement in performing the detrending of individual temperature profiles. See Preusse et al. [2000c] for further details.

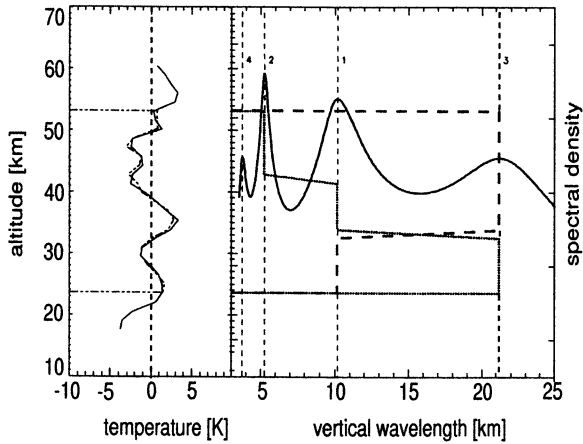


Figure 12: Example of the way in which the MEM/HA analysis method works, in application to CRISTA1 residual temperature data. This profile (solid curve, left panel) was acquired on November 9, 1994 at $\sim 24^\circ\text{S}$ and 82°W over Chile. Right panel shows the MEM spectrum, with the two major peaks highlighted by dotted lines at wavelengths ~ 5 and 10 km. The spectral density scale is logarithmic. Dotted curve on left shows reconstructed profile based on sliding harmonic analysis of the original profile constrained by these two major peaks in the MEM spectrum. We see this fit captures most of the detail in the original residual temperature profile

After we have removed the background, we are left with a global distribution of profiles of residual temperatures $\tilde{T}(x, y, z)$. We search for gravity wave fluctuations in these profiles using a combination of spectral analysis using the Maximum Entropy Method (MEM), and a (vertically) sliding harmonic analysis fit (HA) constrained by the two peak wavenumbers identified in the MEM spectrum. We refer to this hereafter as the MEM/HA analysis method.

The method is best illustrated using an actual example, which is given in Figure 12. The original residual temperature from CRISTA1 data is plotted with the solid curve in the left panel, and its MEM spectrum is shown in the right panel. Four local maxima in the MEM spectrum are evident, with the major two being at wavelengths of ~ 5 km and ~ 10 km. These two harmonics were then used to perform harmonic fits that were slid vertically in 1.5 km vertical increments through the full height range of the profile. A reconstructed profile based on these fits is shown as a dotted curve in the left panel of Figure 12.

The MEM/HA analysis is useful in that it gives a peak vertical temperature amplitude for each harmonic, \hat{T} , at each height level in the profile. It is also easy to automate this analysis, allowing us to analyze and store results for the large number of profiles provided by satellite data. Most of the results in following sections plot estimates of these \hat{T} amplitudes at given levels in the satellite data.

We have compared the MEM/HA analysis method carefully against more standard Fourier

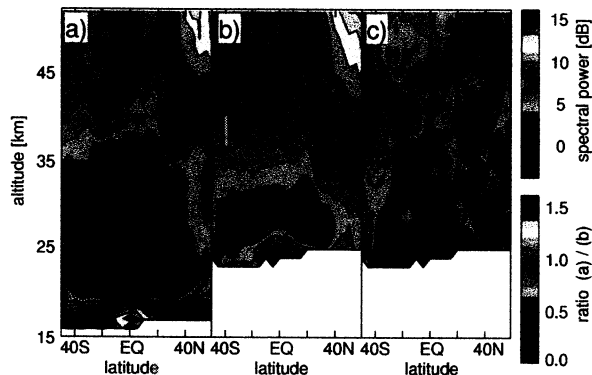


Figure 13: Zonal-mean global \hat{T} amplitudes estimated from CRISTA1 data using (a) MEM/HA and (b) sliding FT analysis. Panel (c) shows the ratio of the two. Since FT requires a finite window in which to form spectral estimates, the height range is less than that for MEM/HA. For more details, see Preusse *et al.* [2000c].

Transform analysis methods that have been used in vertical profile data from radiosondes and lidars. Here we use a sliding Fourier Transform (FT) method to fit the profile in a manner more similar and comparable to output provided by the MEM/HA analysis. Results are summarized in Figure 13, which shows output from MEM/HA analysis (panel a), FT analysis (panel b), and the ratio of the two (panel c). We see from Figure 13 that the MEM/HA method gives better results, particularly at low levels where it captures the tropical peak and subtropical minima in the lower stratosphere better than does the sliding FT analysis (see also Figures 10a and 10c). Note that the LIMS data in Figure 10a were simple standard deviations of Kalman-filtered residuals $\tilde{T}(x, y, z)$.

2.3.2 Stratospheric Mountain Waves

2.3.2.1 CRISTA Data Analysis

Our analysis of mountain waves in CRISTA data has been a major focus of this project, and has been reported extensively elsewhere, in both reports for this project [section 1.1 of report dated December 18, 1999-March 18, 2000; section 1.1 and 1.2 of report dated June 18-September 18, 1999; section 1.1 of report dated March 18-June 18, 1999] and in publications [Eckermann and Preusse, 1999 (see section 7.1); Britt, 1999 (see section 7.3); Eckermann *et al.*, 2000a; Eckermann *et al.*, 2000b; Preusse *et al.*, 2000c]. Thus, we will give an overview of the major results: fuller details can be found in the aforementioned references.

Analysis of residual temperature amplitudes \hat{T} from CRISTA data yields global maps like the one shown in Figure 14 for CRISTA1 temperature data at a height of 25 km, acquired on November 6, 1994. What is immediately evident is a clustering of large amplitude values at the southern end of South America. Given that the Andes Mountains are located here, an immediate question is whether these values here are due to mountain waves generated by the Andes that have propagated into the stratosphere.

Figure 15 summarizes a detailed analysis of a sequence of three successive profiles acquired by one the three CRISTA observing telescopes as it sampled tangent altitudes at regions labeled 1, 2 and 3 in Figure 15c. Large oscillatory temperature amplitudes were evident in these profiles, as shown in the residual data in Figure 15b. That these might be mountain waves is evidenced by the astronaut photograph, shown in Figure 15a, which was taken 2 days earlier as the SPAS was deployed into orbit from the shuttle payload by the shuttle's robot arm. The photograph shows characteristic banded mountain wave clouds downstream of the Andes, which suggests that mountain waves were being produced in the troposphere by flow across the Andes in the days leading up to the stratospheric measurements in Figure 15b.

Inspection of Figure 15b shows that the vertical wavelength of these waves is similar in all three profiles (though phase shifted) at ~6-7 km. A long stationary plane hydrostatic mountain wave emanating from the Andes should have a vertical wavelength given by the hydrostatic relation

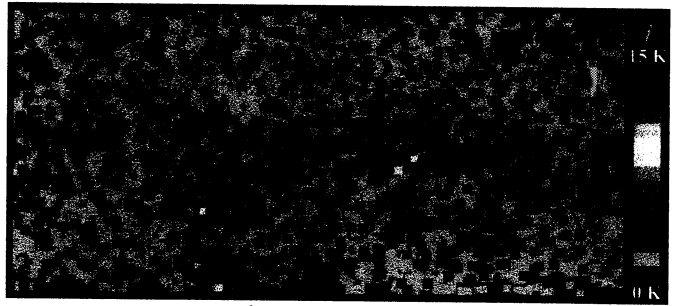


Figure 14: Global \hat{T} amplitudes from CRISTA1 data on November 6, 1994 at 25 km, after correction for visibility effects.

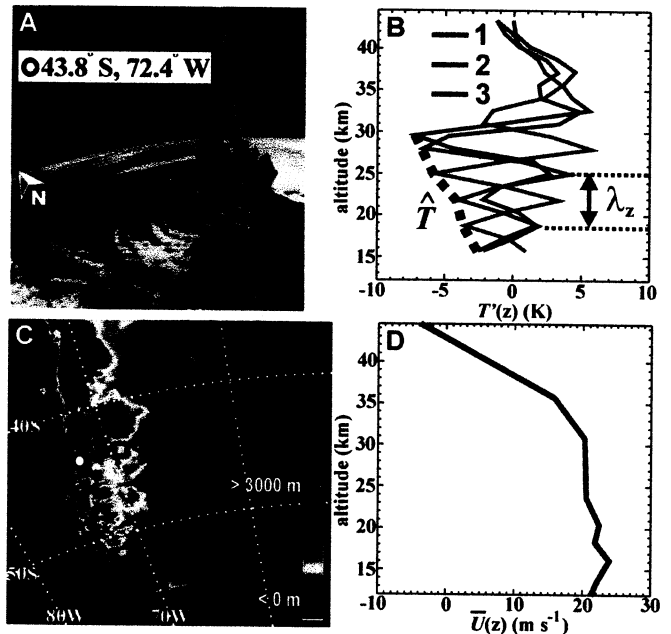


Figure 15: (A) Astronaut photograph over southern South America showing deployment of CRISTA-SPAS from Atlantis, 4th November, 1994, 12:19 UT. The cylindrical assembly on the satellite houses CRISTA. The east coast of Argentina and Atlantic Ocean are visible to the right, banded mountain wave clouds downstream of the Andes are visible to the lower-left. (B) Temperature perturbations $T'(z)$ from three successive CRISTA scans (labeled 1, 2 and 3) on 6th November, 1994 at ~06:00 UT. (C) Topographic elevations for southern South America (linear color scale). Squares labeled 1, 2 and 3 show locations of the profiles in B. The nadir point and orbital motion of the shuttle in A are shown with a black and white circle and black arrow, respectively. (D) Horizontal winds orthogonal to the Andean ridge, from DAO assimilated data [Coy and Swinbank, 1997] at 42°S, 69°W on 6th November, 1994, 06:00 UT.

Estimate of the Effect of Andean Mountain Wave Breaking
on the Background Circulation of the Stratosphere

$$(\lambda_z)_{theory} \approx \frac{2\pi\bar{U}}{N}$$

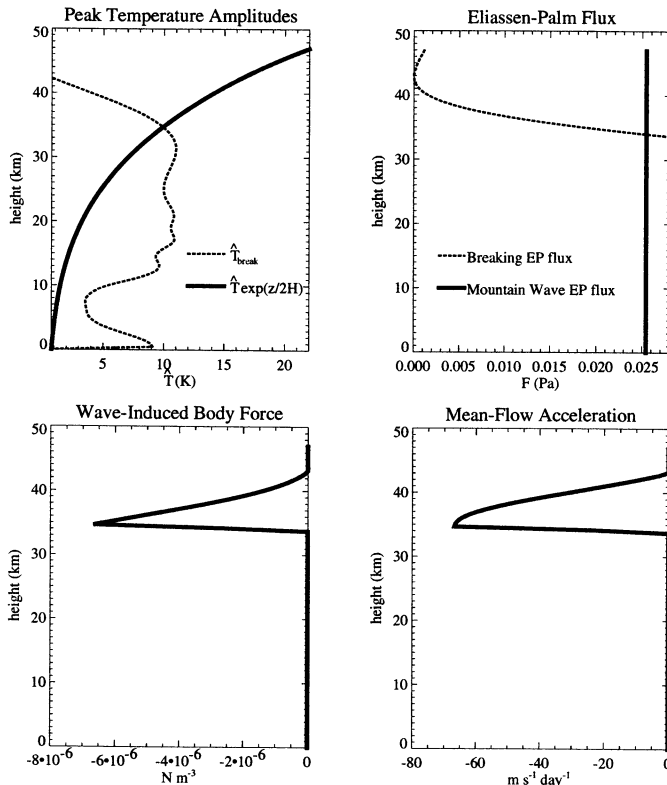


Figure 16: Top row: left panel: thick black curves show the exponential growth of mountain wave temperature amplitude \hat{T} with height, tuned to the 7K amplitude observed at 30 km (Figure 15b). Blue dotted curve shows the calculated breaking amplitude \hat{T}_{break} evaluated from Eq. 2 of Eckermann and Preusse [1999] using the wind profile in Figure 15d. Right panel show corresponding Eliassen-Palm (EP) fluxes, evaluated using the formula derived in citation (17) of Eckermann and Preusse [1999] (see section 7.1). Note that the nonbreaking EP flux is constant with height. Both figures show that wave breaking sets in just above 30 km, where \hat{T} exceeds \hat{T}_{break} .

Lower panels show the ensuing EP-flux divergence (left panel) and the mean-flow acceleration that it implies (right panel). Peak decelerations of $\sim 70 \text{ m s}^{-1} \text{ day}^{-1}$ are noted. These calculations assumed no smearing in the CRISTA measurement, $\lambda_h = 400 \text{ km}$ and $\lambda_z = 6.5 \text{ km}$.

where \bar{U} is the local horizontal wind speed across the Andean Ridge and N is the background buoyancy frequency. At 15-30 km, $\bar{U} \approx 20\text{-}23 \text{ m s}^{-1}$, as can be seen from the assimilated data for this region taken from the Data Assimilation Office ESTRATF analyses for 6 November, 1994 at 06:00Z, plotted in Figure 15d. Since $N \approx 0.020 \text{ rad s}^{-1}$, this yields $(\lambda_z)_{theory} \approx 6\text{-}7 \text{ km}$ according to the mountain wave formula above. The observed vertical wavelength λ_z of the measured wave oscillations in Figure 15b is $\sim 6.5 \text{ km}$ at 15-30 km. The close agreement between observed and theoretical vertical wavelengths is powerful evidence that these are stratospheric mountain waves produced by flow across the Andes.

Successive profiles in Figure 15b are separated horizontally by 200 km, from Figure 15c, and adjacent profiles are phase shifted by 180° (Figure 15b). This implies a horizontal wavelength for this stratospheric mountain wave of $\lambda_h = 400/n \text{ km}$, where n is an odd integer. Given the sensitivity of CRISTA to $\lambda_h > 100\text{-}200 \text{ km}$ (depending on the orientation - see Figures 3,4 and 5), then λ_h must be either 130 or 400 km.

These and other comparisons with linear hydrostatic mountain wave theory confirm the reality of these oscillations as mountain waves – see the *Science* paper of Eckermann and Preusse [1999], reproduced in section

7.1 of this report. In particular, the analysis outlined there shows that the diminution of activity above 30 km in Figure 15b is likely due to strong dissipation of wave activity. Indeed, this diminution can be used to assess the driving this wave breaking has on the background flow. Results of such calculations are plotted in Figure 16 (see report dated June 18-September 18, 1999 for details of these calculations). We see that there is a sharp onset in wave breaking just above 30 km, with values around $50 \text{ m s}^{-1} \text{ day}^{-1}$ in the westward direction (i.e., against the eastward stratospheric flow in Figure 15d). In this linear saturation model calculation, the mountain wave dissipates entirely below 43 km, where a critical level ($\bar{U} = 0$: Figure 15d) causes total absorption of wave activity below these altitudes.

2.3.2.2 Modeling of CRISTA Mountain Wave Signals

To assess whether these and other signals in the CRISTA1 and CRISTA2 temperature data can be explained as mountain waves, we need to go beyond simple analytical formulae and compare the results with more detailed models. During this project, we have conducted an extensive series of modeling experiments to both validate and study the stratospheric mountain waves evident in CRISTA data. Detailed descriptions of this work have been provided in previous reports [sections 1.1 and 1.2 of report dated 18 March-18 June, 2000; section 1.4 of report dated September 18-December 18, 1999; section 1.3 of report dated June 18-September 18, 1999; section 1.2 of report dated March 18-June 18, 1999; section 1.2 of report dated December 18, 1998-March 18, 1999]. Results of some of this work have also been published elsewhere [Eckermann and Preusse, 1999 (see section 7.1); Tan and Eckermann, 2000; Preusse et al., 2000c]. Thus, we will summarize here only the major results and provide a general overview of the three year's worth of work on this topic. Fuller details are to be found in the references above. We present the results in terms of the separate models we have used to study mountain waves in the CRISTA data.

NRL Mountain Wave Forecast Model (NRL/MWFM)

The NRL/MWFM model has been developed in Code 7640 at the Naval Research Laboratory over a number of years. First versions of the model were described by Bacmeister [1993] and Bacmeister et al. [1994]. This version of the model used two-dimensional hydrostatic irrotational wave equations and simple linear saturation criteria, a model we refer to as MWFM 1.0. Since then we have

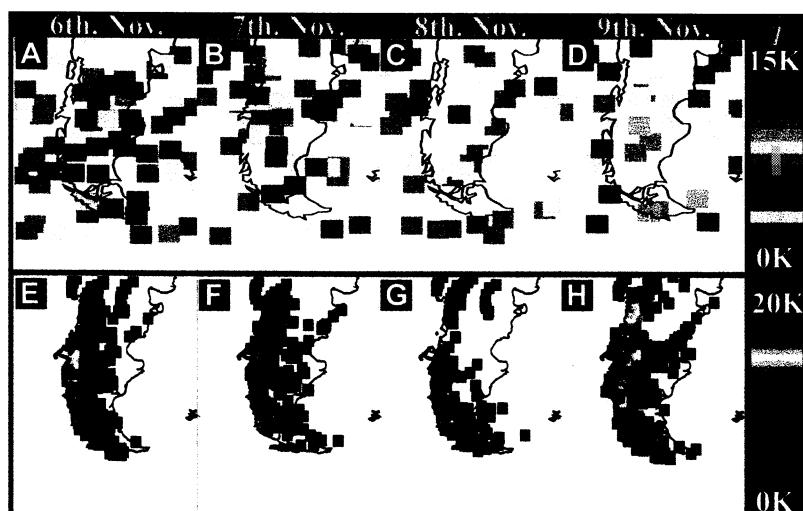


Figure 17: \hat{T} amplitudes over southern South America at an altitude of 25 km from 6th-9th November, 1994 (panels A-D, respectively), plotted as color pixels at the profile's limb tangent location. Raw values were scaled by their vertical wavelength-dependent one-dimensional visibility function (see section 2.2) to correct for observational degradation of amplitudes. Beneath each panel is a corresponding MWFM 2.0 "hindcast" using DAO assimilated winds and temperatures at 12:00 UT for these dates (panels E-H). Modeled mountain waves with $\lambda_z < 5$ km were not plotted since CRISTA is not sensitive to them. The color scales are linear. Pixels are plotted in order of ascending amplitude, and are sized to maximize coverage and visibility while minimizing pixel overlap. Plot after Eckermann and Preusse [1999] – see section 7.1.

undertaken a major project to upgrade the model to use three-dimensional ray-tracing equations to model the three-dimensional radiation of mountain waves governed by a nonhydrostatic rotational dispersion relation. Wave amplitudes are governed by a wave action conservation equation with convective and dynamical saturation criteria under the same approximations. General information on these upgrades is provided by Eckermann and Preusse [1999] and Eckermann et al. [2000a, 2000c], while the theoretical underpinnings of these model extensions were outlined by Broutman et al. [2000]. This new version of the model is known as MWFM 2.0, and was used for the first time in detailed mountain wave forecasting during winter of 1999-2000 in support of NASA's SAGE III Ozone Loss and

Validation Experiment (SOLVE)¹ For further background information on the MWFM model, see <http://uap-www.nrl.navy.mil/dynamics/html/mwfm.html>. All the results shown hereafter were generated using the latest MWFM 2.0 model.

We have used the MWFM in “hindcast mode” to model mountain waves in the CRISTA data. In this mode, we take large-scale analysis winds and temperatures for dates appropriate to the CRISTA mission to “hindcast” mountain wave amplitudes in the stratosphere. These results are compared with global or regional maps of the \hat{T} amplitudes actually measured by CRISTA during these times. In these comparisons, short wavelength mountain waves are removed from the analysis. This is based on the importance of considering such effects to make effective and meaningful comparisons between satellite observations and theory, as proposed by *Alexander* [1998] and verified experimentally in section 2.2.

An example from CRISTA1 is shown in Figure 17. Top row shows \hat{T} amplitudes from CRISTA over November 6-9, 1994, while bottom row shows corresponding MWFM hindcasts using DAO assimilated winds and temperatures at 12Z on these dates. Note that simulated waves with vertical wavelengths < 5 km were removed from the MWFM results, since CRISTA is not sensitive to them (see Figure 3). We see excellent correlations between actual CRISTA data and the MWFM hindcasts. The activity in both is large on the 6th, wanes on the 7th and 8th, then reintensified on the 9th. This is good further evidence that this activity over southern South America is due to mountain waves forced by flow across the Andes. Further evidence

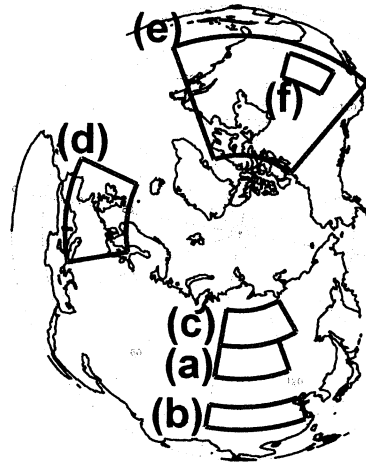


Figure 18: Geographical locations where CRISTA1 data and MWFM hindcasts are analyzed in Figure 19.

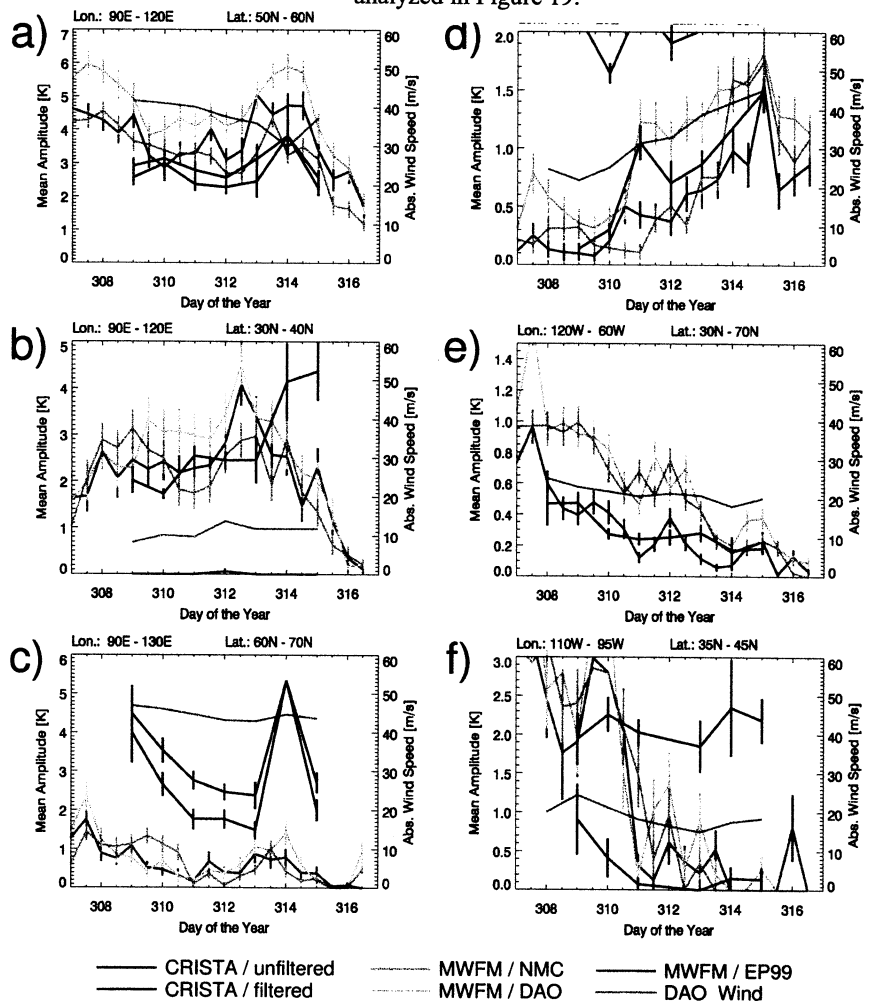


Figure 19: Time series from regions in Figure 18 of filtered (navy curve) and unfiltered (black curve) CRISTA \hat{T} data, and MWFM hindcasts using NCEP analyses (aqua), DAO E0106 analysis (orange), DAO ESTRATF analysis (red). Also plotted are local DAO wind speeds (pink, axis scale to right). Error bars are standard deviations. All plots are for an altitude of 10 hPa

¹ See http://uap-www.nrl.navy.mil/dynamics/html/mwfm_solve1.html

based on accumulated wavelength estimates over the entire CRISTA 1 mission period is provided in section 7.1 by *Eckermann and Preusse* [1999].

We also studied possible mountain wave signals in the northern hemisphere. We studied time series of CRISTA 1 and MWFM hindcasts at six northern hemisphere locations shown in Figure 18, labeled (a) through (f). Time series of \hat{T} data from CRISTA, as well as time series of various MWFM hindcasts, for these regions are shown at an altitude of 10 hPa in Figure 19. The local DAO wind speeds are also plotted with pink in each panel. The comparisons among data and hindcasts in some regions are particularly good, such as geographical regions “a” (central Siberia) and “d” (northwestern Europe),

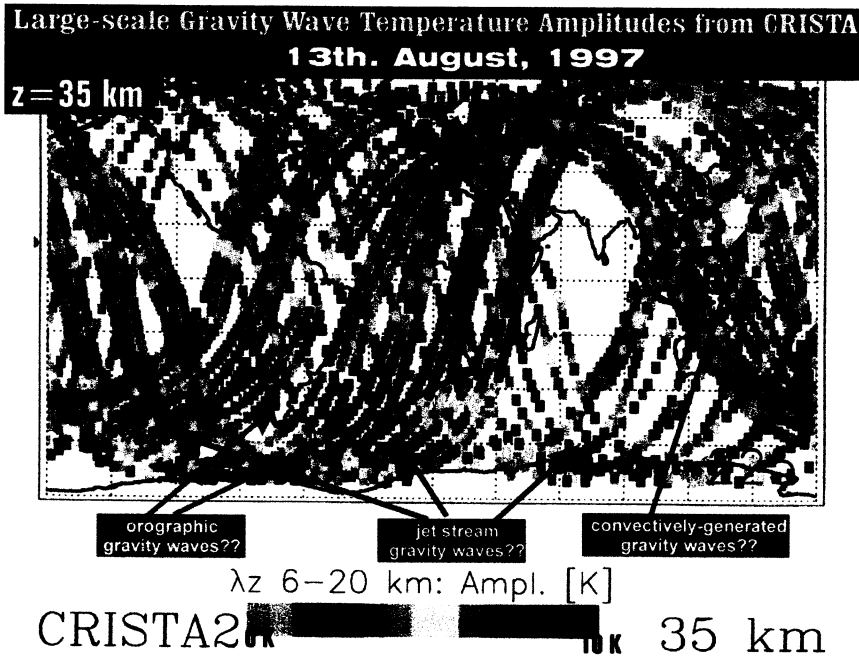


Figure 20: CRISTA peak temperature amplitudes on August 13, 1997 at an altitude of ~35 km. Regions highlighted by aqua box indicate possible zones of stratospheric mountain wave activity. Vertical wavelengths here are in the range 6-20 km.

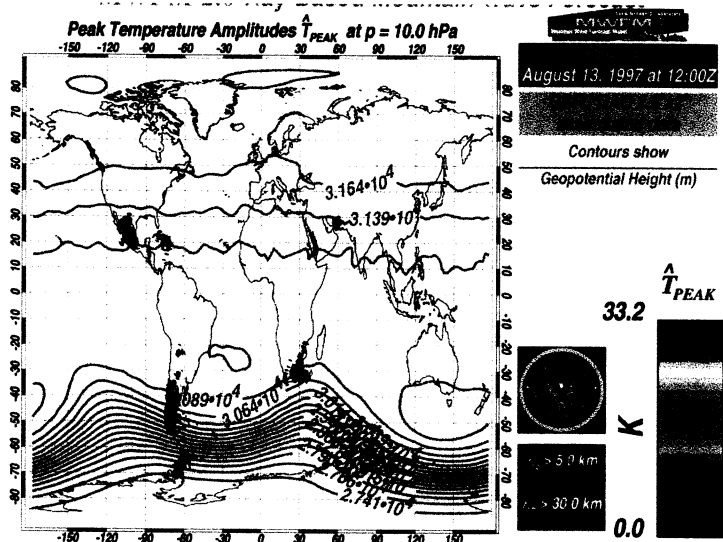


Figure 21: MWFM hindcast for 13 August, 1997 at 12Z, based on DAO assimilated global winds and temperatures for the period. Vertical wavelengths > 5 km and horizontal wavelengths > 30 km only are plotted.

while the comparisons are less compelling elsewhere (e.g., region “c,” northern Siberia). These comparisons lead us to believe that fluctuations in certain broad regions of the northern hemisphere stratosphere during the CRISTA1 mission period are consistent with stratospheric mountain waves.

We have initiated similar studies of fluctuations in data from the CRISTA2 mission during 7-16 August, 1997. Wave activity from this mission showed more activity concentrated in broad regions around the southern polar vortex, as shown in Figure 20. Other regions of activity are analyzed in more depth in later sections. However, here

on August 13, 1997 at 35 km, we note regions over southern South America (as for CRISTA1) and South Africa which may be associated with mountain waves. Results in Figure 20 are plotted within the range of resolved vertical wavelengths between 6 and 20 km.

Figure 21 shows a corresponding NRL/MWFM hindcast simulation at 10hPa for August 13, 1997 at 12Z, based on DAO assimilated global winds and temperatures for this period. Pink contours show geopotential heights, revealing the intense wavenumber-2 vortex structure that existed in the southern hemisphere stratosphere at this time. The hindcasts reveals regions of mountain wave activity over the southern Andes and also over South

Africa, much as observed in Figure 20. Note again here that only longer wavelength waves are retained in these hindcasts, to make more appropriate comparisons with the waves resolved by CRISTA. The presence of mountain waves over southern Africa is associated with the bulging of strong vortex edge winds into this region at the time (evidenced by the closely spaced pink geopotential height contours) via a wavenumber 2 pattern which provides a strong wind channel for mountain waves to reach the stratosphere with long vertical wavelengths (necessary for them to be resolved in the CRISTA2 limb-scanned temperatures). We even see additional evidence for mountain wave activity in the northern hemisphere over eastern Mexico in both the MWFM hindcast map and the CRISTA 2 data. Thus, we have good evidence that mountain waves were evident in the stratosphere and were measured by CRISTA during the CRISTA2 mission period.

Two-Dimensional Mesoscale Model Simulations (NRL/UNSW Model)

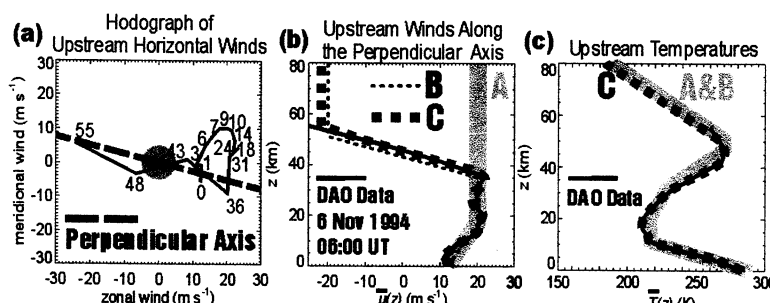


Figure 22: (a) Hodograph of horizontal winds upstream of the Andes from DAO assimilated data on 6th. November, 1994 at 06:00 UT. The altitudes (in kilometers) of each wind value are labeled along the hodograph trace. Orientation of the model experiment axis is shown with broken curve; (b) black solid curve shows upstream DAO wind profile along the perpendicular axis. Other labeled curves show model wind profiles used in Experiments A, B and C using the mesoscale model; (c) vertical profile of upstream DAO temperatures (solid black curve), and the model curves used in Experiments A, B and C.

over the southern Andes in CRISTA1 data (see Figures 14, 15 and 17). The model equations and numerics are set forth in *Tan and Eckermann [2000]*. Using the model, we formulated and performed three experiments, labeled A, B and C, which used wind and temperature profiles as shown in Figure 22. Experiment A used an idealized wind profile that omitted the westward shear layer in the DAO assimilated wind profile above 35 km (Figure 22b). This wind profile is taken from a region upstream of the Andes, shown as the blue “upstream point” in Figure 23. This control experiment allowed us to study mountain wave generation and propagation to stratosphere-mesosphere levels above the Andes without the complication of critical-level effects, and also allowed us to tune the sponge layer in the model. In Experiment B, a linear westward shear layer was introduced above 35 km, yielding a critical level (zero mean wind line) at $z \sim 43$ km. Experiment B allowed us to introduce the critical layer, while maintaining an identical atmospheric situation to Experiment A below 35 km. To this end, both experiments used background temperatures from the 1976 U.S. Standard Atmosphere at mid-latitudes, which fit the upstream DAO temperatures from ~ 0 -55 km quite well (Figure 22c). Experiment C used cubic spline fits to the upstream DAO wind and temperature profiles from 0-55 km. This yielded a zero wind line at $z \sim 46$ km, slightly higher than in Experiment B. Lacking any DAO data to fit above 55 km, from 55-80 km we used constant winds (Figure 22b) and a linear temperature gradient equal to that in Experiments A and B (Figure 22c).

In all three experiments, mountain waves entered the upper stratosphere after a few hours and produced overturning isentropes after ~ 6 hours. Figure 24 plots contours of the total (wave + mean) potential temperature $\Theta(x,z,t)$ after $t = 18$ hours in each experiment over the orographic region of the

While a great deal of work was expended during this project on this particular modeling project, we will only summarize the major final findings of this work, since the full study has been outlined at length in previous quarterly reports (e.g., reports dated 18th September-18th December, 1999, 18th December 1999-18th March, 2000), and has also been published elsewhere in the peer-reviewed literature [*Tan and Eckermann, 2000*]. In short, we developed a new two-dimensional nonhydrostatic compressible mesoscale model and applied it to simulate mountain waves observed

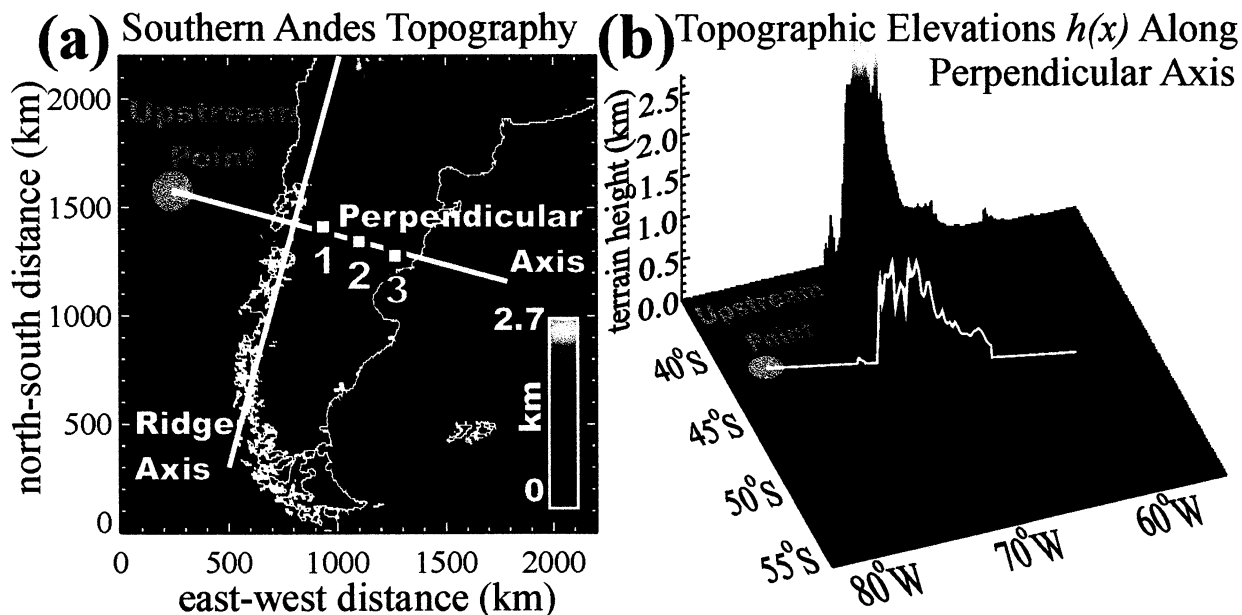


Figure 23: Topographic elevations over southern South America, plotted as (a) filled contours in Cartesian coordinates, (b) three-dimensional elevations in Mercator coordinates. Squares labeled 1,2 and 3 in (a) show locations of temperature profiles acquired by CRISTA. Raw elevations $h(x)$ along the perpendicular axis in (a) are plotted in (b), and represent the topography used in the model simulations. The blue dot in both figures is the upstream point where mean winds and temperatures are computed from DAO data in Figure 22.

Andes. Colors show contours of temperature perturbations $T'(x,z,t)$. We see that the critical levels in Experiments B and C efficiently absorb mountain waves, and yield rapid wave shortening and intense breaking at closely underlying altitudes. Despite the different wind profiles and the significant amounts of overturning (nonlinearity) and mixing by this time, the large-scale wave structures below 35 km look very similar in all three experiments.

The wave fields evolve significantly with time, as illustrated in Figure 25 using results from Experiment C. The middle atmosphere is dominated initially by short horizontal wavelengths, which trigger vigorous overturning and mixing by ~ 8 hours. The turbulent zones persist and seem to move downstream after 10 hours. Thereafter, progressively longer horizontal wavelengths come to dominate

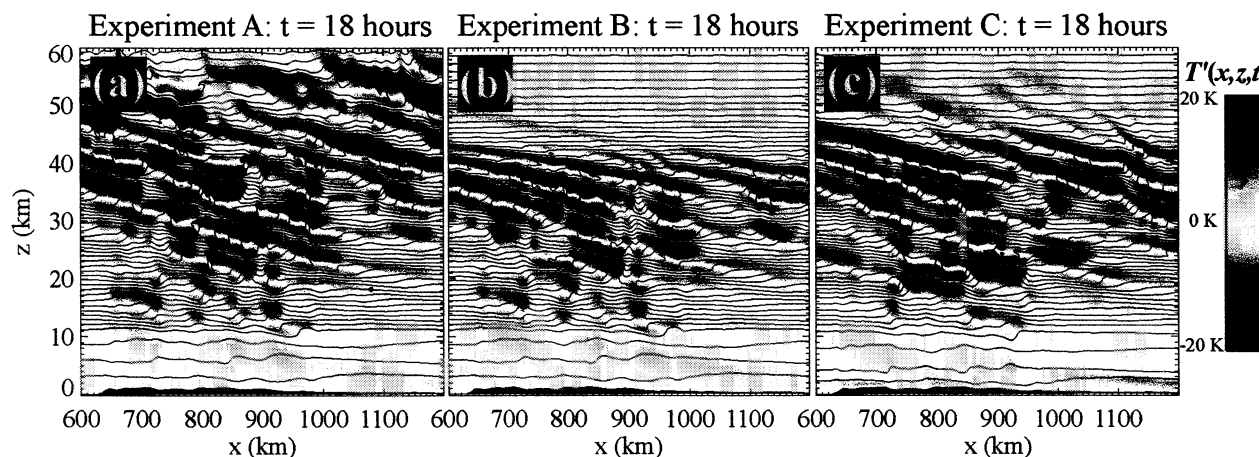


Figure 24: Temperature fields 18 hours into Experiments A, B and C. Contours show isentropes $\Theta(x,z,t)$, with constant logarithmic separation between adjacent contours. Temperature perturbations $T'(x,z,t)$ are overlaid using the blue-red color scheme shown to the right. Underlying topography is shown in black.

the wave field, and less vigorous overturning is evident. After 24 hours, waves at lower stratospheric levels are suppressed compared to their earlier intensities.

The wavelength selection and the resultant group velocities explain the wavelength selection observed here in Figure 24, as well as the time evolution of dominant horizontal wavelengths evident in Figure 25. These theoretical arguments were set forth in previous reports, as well as by *Tan and*

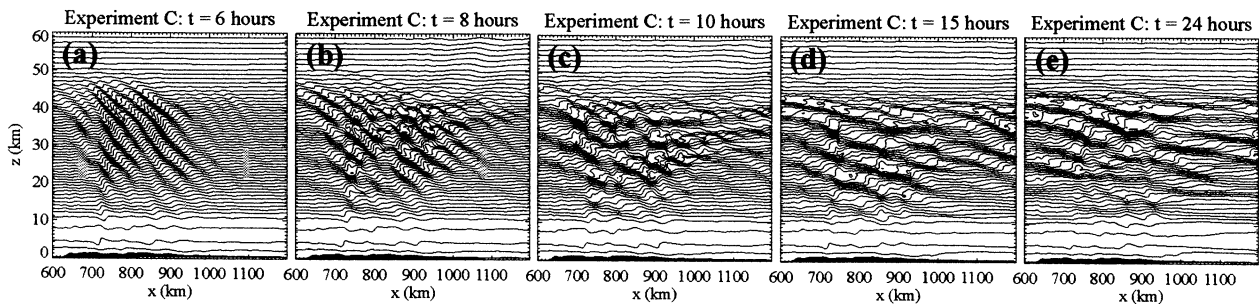


Figure 25: Contours of isentropes $\Theta(x,z,t)$ from Experiment C after $t = 6, 8, 10, 15$ and 24 hours. Same constant logarithmic separation between contours is used in each panel. Underlying topography is shaded.

Eckermann [2000], and will not be repeated here. The simulations out to 24 hours in Figure 25e show much longer horizontal wavelength waves in the stratosphere than earlier, consistent with the long wavelength waves that appear to dominate the response and need to be present in order for the CRISTA and MLS limb scans to resolve them.

The general form of the profiles over the Andes compares favorably with the CRISTA data. Figure 26 plots a sample temperature perturbation profile from Experiment C at $x = 845$ km, a location between points 1 and 3 over the Andes where CRISTA acquired these profiles, as shown in Figure 23a. The CRISTA profiles at these points are overlaid in Figure 26, taken from Figure 15b. The wavelength, amplitude and general height variation of the model profile are all quite similar to the data. Oscillations in Experiment C seem to penetrate slightly higher prior to shortening and dissipating below the critical layer. Note too that the amplitudes of the CRISTA profiles are probably underestimated due to instrumental effects, and that CRISTA cannot resolve any waves of vertical wavelength less than $\sim 3\text{--}5$ km [*Eckermann and Preusse, 1999*; see Figure 3].

These results, which are not tuned in any way but allowed to run from “first principles” using a 2D nonlinear mesoscale model constrained to background conditions at this time, generate mountain waves with many similarities to those measured by CRISTA on 6 November, 1994. This provides powerful evidence that CRISTA indeed measured long wavelength mountain waves over the Andes at this time.

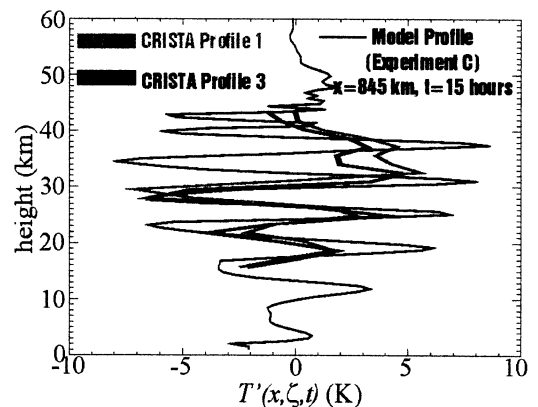


Figure 26: Black curve shows a vertical profile of temperature perturbations at $x = 845$ km, 15 hours into Experiment C. Two shaded solid curves reproduce the temperature perturbation profiles acquired by CRISTA at locations 1 and 3 in Figures 15a, 15b and 23a.

Three-Dimensional Mesoscale Model Simulations (MM5)

In collaboration with Dr. Andreas Dörnbrack of the Deutsche Institut für Luft und Raumfahrt (DLR) in Oberpfaffenhofen, Germany, we have also used a version of the Penn State Mesoscale Model 5 (MM5) to provide hindcasts of large-scale mountain waves for the CRISTA1 mission, in a similar way to what has been done previously in the stratosphere over northern Scandinavia [e.g., *Dörnbrack et al., 1998*] and Greenland [e.g., *Leutbecher and Volkert, 2000*]. Given the MM5 model’s ability to simulate

the largest scale mountain waves, which CRISTA and MLS measure, then we began discussions about performing MM5 simulations over southern Scandinavia during the CRISTA 1 mission period to see if this model could cast light on the three-dimensional nature of mountain waves radiated into the stratosphere from this region.

In these simulations, the MM5 model was initialized using archived analysis ECMWF winds and temperatures for 5th November, 1994 at 0:00 Z, and then was integrated for 36 hours thereafter in dry mode with an upper level of ~28 km and a horizontal grid resolution of ~27 km. The temperature results are shown in Figure 27c at an altitude of 30 hPa. Another nested grid simulation was conducted which was identical to the one in Figure 27c, but increased the effective horizontal resolution. Corresponding results from that simulation are shown in Figure 27d. As for the 2D mesoscale simulations, we see a long stratospheric mountain wave emerge in the MM5 results, of a horizontal scale long enough to be resolved by CRISTA (see Figure 3). Interestingly, however, these waves in the MM5 output are titled somewhat in phase with respect to the Andean topography, as evident from inspecting Figures 23 and 27c,d. The reason for this may be due to surface wind patterns that bulge southward from the Andes and then northward past the Andes in response to the larger Andean obstacle heights to the north [see, e.g., *Seluchi et al.*, 1998]. Another reason could be the sharp latitudinal shear in the vortex winds here that may refract waves latitudinally. It might also be caused by flow across the “hook”-like topography near the southernmost tip of South America (see Figure 23b). However, since these orientations persist even to the north, the exact reason for this in the MM5 simulations is unclear and is a somewhat unexpected finding.

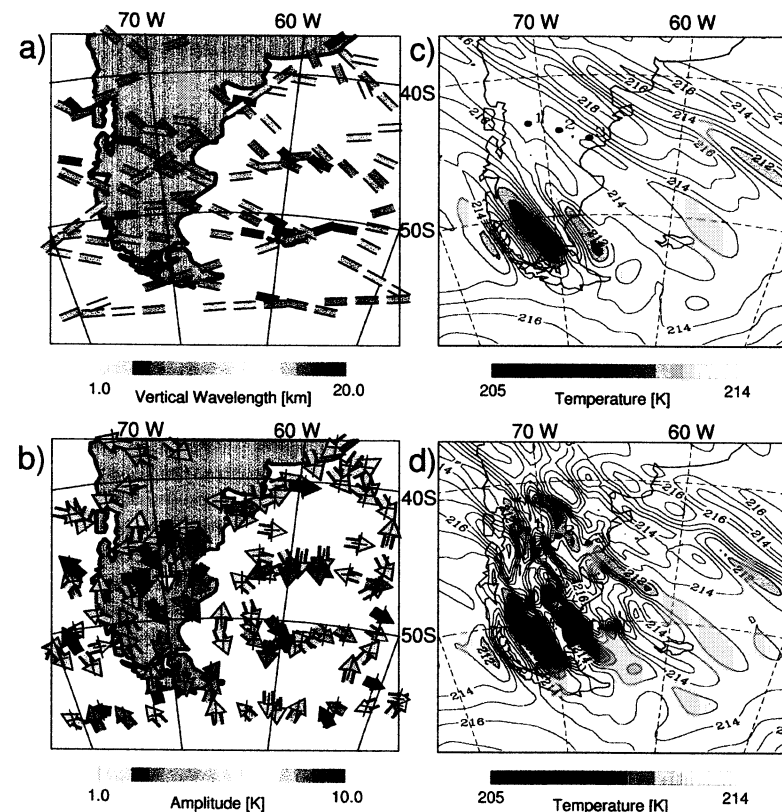


Figure 27: Gravity wave activity over southern South America at 30 hPa. The vertical wavelength in the CRISTA1 profile and the viewing direction of the instrument are displayed in (a). Panel (b) shows CRISTA wave amplitudes (color scale), along with wave phase depicted by the orientation of the arrow. Panel (c) shows contours of temperature at 30 hPa 36 hours into the MM5 simulation for this period, while Panel (d) shows results using a nested grid simulation to increase the horizontal resolution to ~14 km. Red dots show locations of the 3 CRISTA1 profiles in Figures 15b, 15c and 23a.

Activity at 30 hPa from CRISTA1 data at this time are profiled in a more in-depth way in Figures 27a and 27b, showing wavelength, phase, amplitude and telescope limb scan directions, all of which aid in assessing how fluctuations like those seen in the MM5 simulations might be resolved by the CRISTA1 limb scan data. We see that the interesting 45° tilt of the MM5 wave fronts with respect to the Andean Ridge axis in Figures 27c and 27d assists in CRISTA resolving the large waves noted in the three profiles noted in Figures 15, 23a and 26, since, as noted from Figure 27a and 27b, the telescopes observed these waves roughly along a 45° orientation that runs roughly parallel to a wave phase front in Figure 27c and 27d, a situation akin to the “best case” observing scenario in Figure 4. The absolute magnitudes of the temperature amplitudes are similar, with MM5 values generally a little lower. We also see evidence of significant downstream penetration

of mountain wave activity in both the CRISTA1 data and the MM5 hindcast simulation. However, the major difference is that the activity in the MM5 simulations is concentrated near the southernmost tip of South America, whereas in the CRISTA data for November 6, the largest burst is further north. Nonetheless, the comparisons are generally quite good, and add further weight to our accumulating body of evidence that these fluctuations in the CRISTA1 temperature channel over the southern Andes are due to long wavelength stratospheric mountain waves. Initial results of this work are being written up as part of a major paper detailing the mountain wave analysis and modeling from the CRISTA1 mission [Preusse *et al.*, 2000c].

NCEP Global Analyses

Around 1995, the National Centers for Environmental Prediction (NCEP) global upgraded the output from their global spectral forecast model to a $1^\circ \times 1^\circ$ resolution. Accordingly, NCEP started issuing $1^\circ \times 1^\circ$ analysis fields in support of these new forecasts, in addition to the standard $5^\circ \times 2^\circ$ fields that have been issued for many years, as well as the $2.5^\circ \times 2.5^\circ$ fields that are being issued as part of the NCEP 40-year reanalysis project [Kalnay *et al.*, 1996]. At mid-latitudes, 1° longitudinal resolution corresponding to a spatial resolution $\Delta x \sim 75$ km, and thus, given that CRISTA1 cannot resolve waves much shorter than 200 km, this raises the interesting possibility that waves evident in the CRISTA temperature channel might also be evident in these new high-resolution NCEP global analyses. Since the CRISTA1 mission took place in November, 1994, just before these analyses were issued, we cannot test this for the CRISTA1 mission, but we can test these ideas during the CRISTA2 mission period of August, 1997.

The exact analysis methods we used are discussed at length in the following section detailing potential jet stream-related activity in the CRISTA2 data. Suffice to say here that we analyze divergences D in the NCEP analyses, which often show gravity wave fluctuations directly [e.g., O'Sullivan and Dunkerton, 1995]. Figure 28 shows a longitude-height section along the Andes at a latitude of 49° S on August 12, 1997. We see coherent wave fluctuations in the divergences D with phase lines that slope upward and westward,

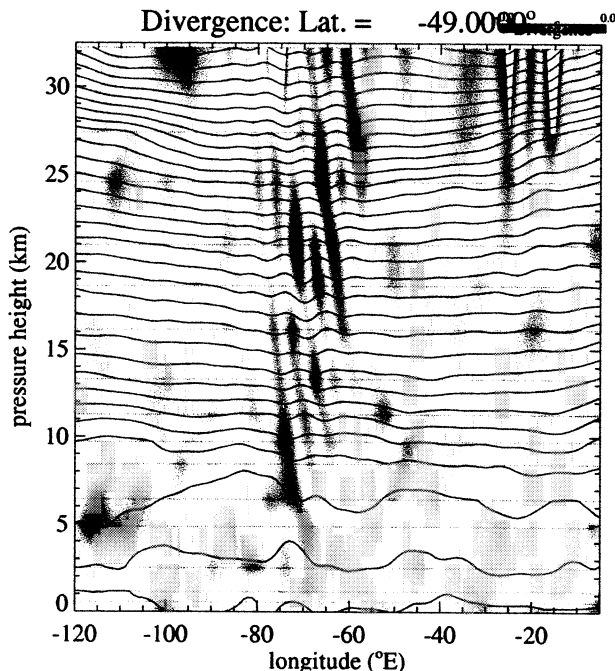


Figure 28: Longitude-height section of divergence D calculated from $1^\circ \times 1^\circ$ NCEP global analyses for August 12, 1997 along a latitude of 49° S. In the longitude range $70^\circ - 80^\circ$ W, corresponding to the southern Andes, we see wave fluctuations with phase lines that slope upwards and westward, consistent with a mountain wave propagating westward with respect to the eastward background winds. Note the increase in vertical wavelength as the wave propagates into the strong eastward vortex winds in the stratosphere. Black contours show isentropes, which also reveal perturbations due to these waves.

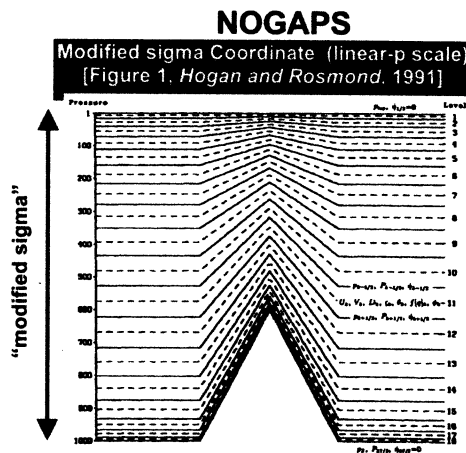


Figure 29: Illustration of how the sigma coordinate transformation over topography extends to the top model altitude (based on NOGAPS model of Hogan and Rosmond [1991]). The pressure scale here in linear, which accentuates the obstacle effect

2.3.3 Stratospheric “Vortex Gravity Waves”

Figure 30 shows a time series of four successive days of peak amplitudes \hat{T} of small-scale temperature fluctuations due to gravity waves in CRISTA2 data, isolated as before using the MEM/HA analysis method applied to Kalman-detrended global temperature fields for each day of the mission. As noted in Figures 20, 21 and 28, activity over the southernmost regions of South America and Africa appears to be due to long stratospheric mountain waves. However, unlike the CRISTA1 mission which did not show a great deal of other activity at these southern latitudes (see, e.g., Figure 14), the stratospheric activity during CRISTA2 in Figure 30 shows broad regions of activity at southern mid-latitudes that exist well away from significant topography or even land mass. These mysterious “vortex gravity waves” are the most notable feature of the Southern Hemisphere CRISTA2 gravity wave data in Figure 30, and require dedicated study.

Similar enhancements of stratospheric gravity wave activity in and around the polar vortices have been inferred from MLS data [Wu and Waters, 1997; Jiang and Wu, 2000], as well as LIMS data [Fetzer and Gille, 1994] and data from ground-based lidar instruments [e.g., Whiteway, 1999; Duck and Whiteway, 2000]. The major source(s) of these waves remain(s) undetermined. In the northern hemisphere (Arctic), it has been argued that wave variance enhancements that correlate with intensifying vortex flow overhead are due to increased transmission of mountain waves into the stratosphere [e.g., Whiteway et al., 1997; Whiteway, 1999], although correlations with tropospheric jets have also been noted [Duck and Whiteway, 2000]. Since most of the southern hemisphere vortex activity in Figure 30 is over open ocean, these southern hemisphere features cannot be explained by mountain waves alone (compare Figures 21 and 30). Thus, we must consider other sources in trying to explain these waves in the CRISTA2 data.

We have considered unstable jet stream dynamics as a potentially viable source of the activity in around the southern vortex that we see in the CRISTA 2 data in Figures 20 and 30. This is because the region at $\sim 40\text{--}50^\circ\text{S}$ where stratospheric vortex winds peak is also characterized at lower levels by strong jet stream winds and evolving baroclinic wave systems, storms and fronts (the so-called “roaring 40’s” region). Guest et al. [2000] showed that large-scale gravity waves observed in radiosonde data over Macquarie Island (54°S , 159°E) were probably due to inertia gravity waves radiated from jet exit regions, rather than by flow across the island’s orography.

Modeling or identifying global jet-related wave sources is much more difficult than it is for mountain waves: indeed, it is probably fair to say that the theory of waves generated from jet streams is still not fully developed [see, e.g., Vadas and Fritts, 2001]. However, Uccellini and Koch [1987] noted a reproducible synoptic setting for long wavelength gravity waves radiated from the jet: the situation appropriate to the southern hemisphere is shown in Figure 31. Winds in the hatched regions become highly ageostrophic and decelerate, radiating inertia gravity waves in response. Such waves have been observed in idealized global spectral model simulations [e.g., O’Sullivan and Dunkerton, 1995] as well as in data and numerical weather forecasts [e.g., Koch and O’Handley, 1997]. Within spectral numerical weather prediction model models, it is usual to deal with prognostic equations for the horizontal velocity divergence $D = \vec{\nabla}_h \cdot \vec{V}$, where $\vec{V} = (u, v)$ is the horizontal velocity. The prognostic time tendency equation, in pressure coordinates, then becomes

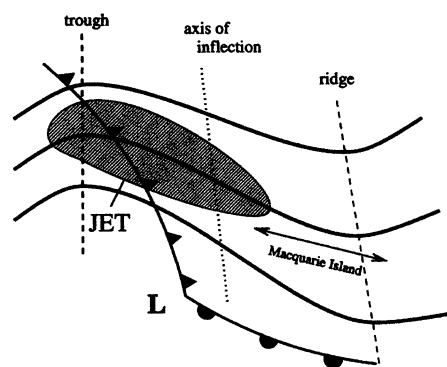


Figure 31: Synoptic setting at 300 hPa for gravity wave radiation by the jet, transposed for southern hemisphere [after Guest et al., 2000]. Downstream of a trough, as height contours spread apart and air parcels decelerate, winds become strongly ageostrophic in the shaded region, and radiate long low-frequency gravity waves as the jet region relaxes to a state of balance.

$$\frac{dD}{dt} = -D^2 - \frac{\partial \vec{V}}{\partial p} \cdot \vec{\nabla}_h \omega + \dot{D}_{\text{nonlinear}}$$

where

$$\dot{D}_{\text{nonlinear}} = 2J(u, v) - \beta u + f\zeta - \nabla_h^2 \Phi$$

[e.g., *Moore and Abeling*, 1988]. Here $d/dt = \partial/\partial t + \vec{V} \cdot \vec{\nabla}_h + \omega \partial/\partial p$ is the total advective time derivative, ω is the vertical velocity (transformed to the pressure coordinate), J is the Jacobian operator, such that $J(u, v) = (a \cos \phi)^{-1} [\partial u / \partial \lambda \partial v / \partial \phi - \partial v / \partial \lambda \partial u / \partial \phi]$, ζ is the relative vorticity, f is the Coriolis parameter, $\beta = \partial f / \partial y$ and Φ is the geopotential. For stable or “balanced” jet stream dynamics, the 4 nonlinear tendency terms in $\dot{D}_{\text{nonlinear}}$ above sum to zero to a very good approximation [*Moore and Abeling*, 1988] and this is sometimes referred to as the nonlinear balance equation, or NBE for short. In jet exit regions like Figure 31, however, $\dot{D}_{\text{nonlinear}}$ can be significantly nonzero, and thus can serve as an identifier of unstable regions where long mesoscale gravity waves may be radiated by unbalanced jet stream dynamics. Indeed, *Koch and O’Handley* [1997] suggested such a calculation as a way to forecast trapped mesoscale gravity waves that can produce locally severe weather. In other words, $\dot{D}_{\text{nonlinear}}$ may serve as a large-scale proxy for identifying synoptic regions where smaller-scale gravity waves radiate from the jet.

To investigate this, we use high-resolution global analyses issued by the National Centers for Environmental Prediction (NCEP). *Kalnay et al.* [1996] provide background on the NCEP reanalyses in the context of their specific 40-year reanalysis project. However, those reanalysis products are issued on a $2.5^\circ \times 2.5^\circ$ global grid, whereas we need better resolution than this to resolve long wavelength gravity waves, as shown for our analysis of mountain waves in Figure 28 and the accompanying discussion. Thus, we again focus on the $1^\circ \times 1^\circ$ analysis product that has been issued since ~1995 in support of the current $1^\circ \times 1^\circ$ global numerical weather predictions from the NCEP global spectral model. These products are issued once daily (at 0Z) on 22 pressure levels from sea level (1000 hPa) up to 10 hPa.

We have studied these NCEP products throughout the second CRISTA-SPAS mission (7-16 August, 1997). First, we computed divergences D directly from these pressure-gridded global NCEP datasets. These estimates should be quite reliable since the NCEP global spectral

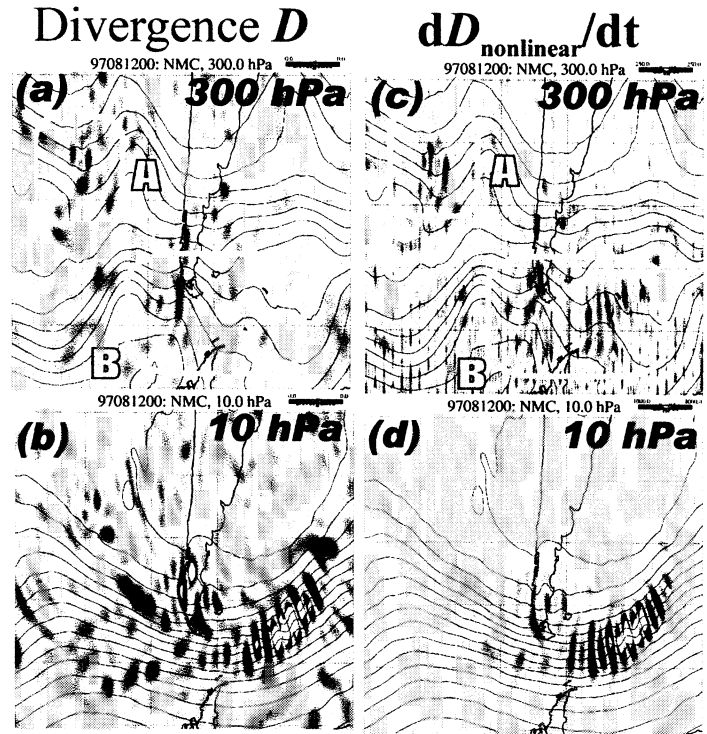


Figure 32: Divergences D (left column) and the nonlinear divergence tendency term, $\dot{D}_{\text{nonlinear}}$ evaluated at 300 hPa (top row) and 10 hPa (bottom row) using data from $1^\circ \times 1^\circ$ NCEP reanalyses for August 12, 1997 at 0Z. The red-blue color scale is shown top-right, with white showing zero, red most positive and blue the most negative values. Largest divergences are 10^{-6} s^{-1} and $\dot{D}_{\text{nonlinear}}$ values are capped at $2.5 \times 10^{-8} \text{ s}^{-2}$ in panel c and $10 \times 10^{-8} \text{ s}^{-2}$ in panel (d). Black lines show geopotential height contours. Yellow regions show synoptic settings resembling Figure ZP, which yield apparent propagating gravity wave disturbances.

model is formulated in terms of equations that solve directly for divergence and vorticity. Furthermore, divergences effectively suppress the mean state, and so can often show directly any coherent mesoscale wave disturbances radiated from jets or other regions [e.g., *O’Sullivan and Dunkerton, 1995*]. We also computed the nonlinear divergence tendency term $\dot{D}_{nonlinear}$ using the four terms based on its defining NBE equation above, to search for unstable jet exit regions that may also radiate propagating wave disturbances [e.g., *Koch and O’Handley, 1997*].

Some results for 12 August, 1997 in the western sector of the globe centered near South America are plotted in Figure 32. The top row of plots shows results at 300 hPa, the nominal jet stream altitude at which previous studies have diagnosed the flow [e.g., *Uccellini and Koch, 1987*]. Divergences D are shown in Figure 32a, while the nonlinear divergence tendency $\dot{D}_{nonlinear}$ is shown in Figure 32c: both are contoured using the blue-red color scheme, with blue showing negative anomalies and red positive ones in each case. Black contours show geopotential height contours.

The yellow squares in Figure 32a,c highlight two regions where the synoptic situation shown in Figure 31 appears to have been established, labeled Cases A and B. We see both D and $\dot{D}_{nonlinear}$ anomalies in these regions that take the form of radiating wavelike disturbances. In the “Case A” region, fluctuations in D and $\dot{D}_{nonlinear}$ appear to radiate away from the jet exit region in Figure 31a,c as a somewhat spherical wave disturbance, consistent with gravity wave radiation from a compact three-dimensional region of instability. In the lower “Case B” region, just below the tip of South America, even more coherent wavelike motions are seen, though here the wave fronts appear more planar.

The bottom row of Figure 32 shows corresponding results at the highest stratospheric altitude of 10 hPa (~32 km). In “Case A,” we see little evidence for these waves in the stratosphere, due presumably to less favorable wind conditions for this activity to penetrate to higher altitudes. For “Case B,” however, we see large amounts of activity of similar horizontal wavelength at 10 hPa related to the activity noted at 300 hPa. Note in particular the strong clustering of geopotential height contours associated with the strong stratospheric winds at the edge of the polar vortex. This is particularly clear in Figure 32d. Interestingly, these strong stratospheric wave trains appear to follow the peak winds of the vortex. This may be due to ducting of these waves into the peak winds of the vortex by strong latitudinal wind shear. For a horizontal wavenumber vector (k,l) , the refraction equation for the meridional wavenumber is

$$\frac{dl}{dt} = -k \frac{\partial \bar{U}}{\partial y}$$

which can yield strong focusing into peak winds with strong latitudinal gradients. For example, strong refractive focusing of waves into the peak

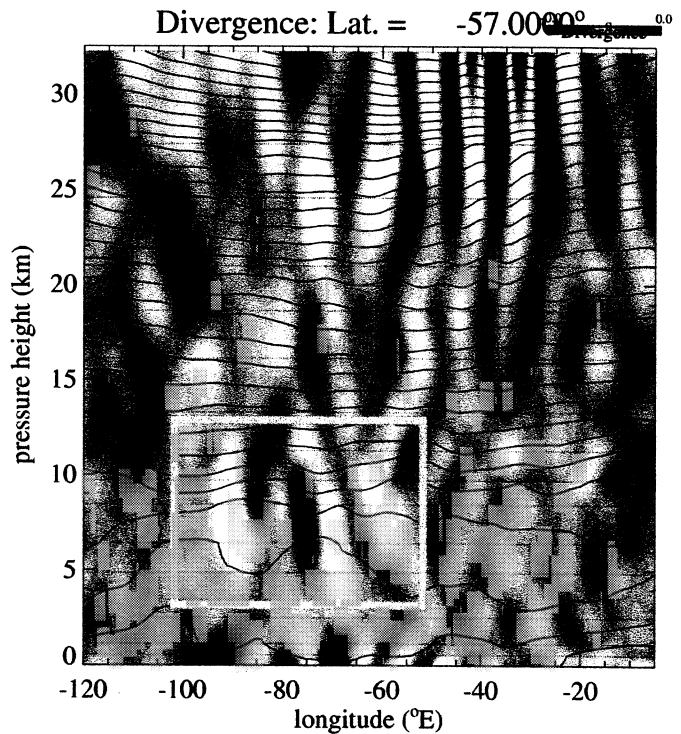


Figure 33: Longitude-height section of divergence D at 57°S for NCEP analyses on 12 August, 1997. Contours show (dry) isentropic levels (logarithmically equispaced). Orange box shows jet exit region with elevated isentropes that descend and spread out downstream. Sloping longitude-height phase contours in D are evident in this region, suggestive of waves produced by this unstable jet region. Note too the intense wave activity of similar horizontal wavelength in the uppermost stratospheric levels, visible in both the divergences and potential temperature contours.

winds of the polar night jet was noted by *Dunkerton* [1984] and *Marks and Eckermann* [1995]. Similar refractive effects nearer the source at jet stream level were noted by *O'Sullivan and Dunkerton* [1995].

To study wave generation in "Case B" in a little more depth, Figure 33 shows a longitude-height section of divergence (blue-red colors) and isentropic surfaces (black contours) at a latitude of 57°S based on the NCEP analyses in Figure 31. Unlike a similar plot at 49°S in Figure 28, this latitude region is well removed from topography of the Andes or the Antarctic Peninsula, and thus waves here cannot be due to mountains. The orange box in Figure 33 shows the region where waves are apparently being generated in Case "B." We see a coherent packet of waves at ~7-12 km and ~90°-60°W, with sloping time-height contours consistent with a wave propagating westward with respect to the background flow. We also see large amplitude waves in the stratosphere above and downstream of this region, though here the phase lines are much deeper due to the intensifying stratospheric flow, which dilates the vertical wavelength of waves aligned along the vortex flow. These waves are evident not just in the colored divergence contours, but also as perturbations on the isentropes in Figure 33.

Nonetheless, the exact correspondence between the low-level activity produced by the jet and the intensifying activity in the stratosphere in Figure 33 is not entirely clear. In particular, unlike with the mountain waves in Figure 28, it is difficult at present to glean clear propagation paths all the way into the stratosphere for this jet-related activity in the NCEP analyses. Analysis of similar maps in NCEP data from other days during the CRISTA 2 missions shows similar ambiguities. We believe this has much to do with the background wind patterns, which may tend to refract waves in and out of the resolved field of view in various cases. In particular, just as for CRISTA data, the vertical wavelength must be sufficiently large that it can be resolved in the final issued data. It may also be that some of the activity at upper levels may be going unstable, due either to wave breaking or due to numerical instabilities, since the mean and wave-induced shears here are large. Indeed, initial computation of Richardson numbers here reveal somewhat low values. We have also analyzed the source regions using other measures of instability or imbalance, such as Richardson number and the Lagrangian Rossby number, the latter based on the approximate relation recommended by *Koch and O'Handley* [1997]. These results also show interesting trends but are not shown here in order to keep this report to a manageable length. This work, much of it conducted in the last 3 monthly quarter, will be a focus of ongoing and future research.

2.3.4 Convectively-Generated Stratospheric Gravity Waves

This work has been reported at various stages in previous quarterly reports (e.g., section 1.1. of report dated June 18-September 18, 1998; section 1.2 of report dated 18th. September-18th. December, 1999). Here we briefly report on consolidated research on this topic. Our research here has focused mainly on data acquired during CRISTA2. During this mission period, a large typhoon developed over the South China Sea, reaching Category 5 ("super typhoon") status on 12 August, 1997. Named "Winnie," it was visible both from the GOES geostationary weather satellite and was also photographed by the shuttle astronauts – images of the typhoon from both orbital platforms are shown in Figure 34.

Stratospheric fluctuations were also evident in CRISTA data in this region north of Indonesia, as can be seen in Figures 20 and 30. Typhoons have been implicated previously as sources of large-scale gravity waves [*Matsumoto and Okamura*, 1985; *Sato*, 1993], so it is interesting to investigate whether this enhanced gravity wave activity is due to convection generated within this typhoon region, since, if so, it would comprise an unusually clear connection between a convective source in the tropical troposphere and enhanced gravity wave activity in the tropical stratosphere. Unfortunately, during the mission a special "hawkeye mode" was used to maximize sampling over the Indonesia area in an effort to search for convectively generated wave activity over the maritime continent. This increased sampling came at the expense of coverage of regions to the north where Super Typhoon Winnie formed and passed.



Figure 34: (a) full disk GOES 8 visible image on 12th August, 1997, showing Super Typhoon Winnie. Australia is visible to the south (image courtesy of NOAA's National Climatic Data Center); (b) close-up photo of Winnie taken from *Discovery* on 13th August, 1997 (shuttle robot arm visible in foreground).

Water vapor concentrations were one among many trace constituents measured by CRISTA during the mission, and these measurements extended to quite low levels. Thus, we can study tropical water vapor distributions down to ~13 km, which is well within the tropical troposphere, and correlate it with stratospheric gravity wave activity. Such an analysis is presented in Figure 35, where each panel is divided into 8 subpanels and shows CRISTA limb-scanned data on successive days over the South China Sea area, starting at Day 220 (August 8, 1997) and finishing at Day 227 (August 15, 1997). Note that the most data intensive days are Days 222-226. The passage of Super Typhoon Winnie over these days is shown in the first panel of Figure 35c. Figure 35a shows the evolution of measured gravity wave activity at an altitude of ~35 km for waves with vertical wavelengths in the 14-30 km range, while Figure 35b shows the same thing for waves in the 9-14 km vertical wavelength range. Figure 35c shows measured water vapor mixing ratios at $z=13$ km as inferred from the CRISTA data, while Figure 35d plots the inferred altitude of cloud tops from these data.

We see what looks to be fairly uniform cloudiness and convection in the area that are

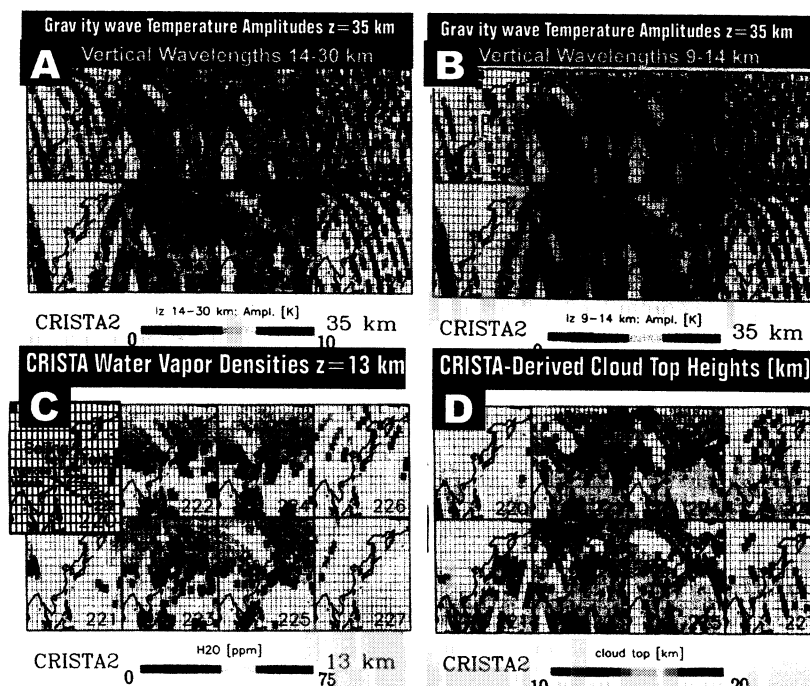


Figure 35: Time sequence of daily CRISTA 2 measurements over China Sea region from Days 220-227 (7-14 August, 1997). **A:** temperature fluctuation amplitudes at $z=35$ km for $\lambda_z = 14-30$ km. **B:** temperature fluctuation amplitudes at $z=35$ km for $\lambda_z = 9-14$ km. **C:** water vapor mixing ratios (ppmv) at $z=13$ km. Storm track of Super Typhoon Winnie as a function of day number is plotted in the top-left panel. **D:** cloud top heights (in kilometers).

well correlated with increases in the gravity wave variance locally. There does not appear to be any obvious correlation between the daily evolution of the stratospheric gravity wave activity measured by CRISTA (Figure 35a,b) and the passage of the typhoon into the South China Sea region and onto the coast of China. Exact interpretation of the convective activity may be difficult too, since the values are taken at $z=13$ km and the abrupt change on moving to more poleward latitudes may be due to passage from the subtropical upper troposphere into the dryer mid-latitude lower stratosphere.

A useful way to diagnose the nature of the wave activity, pioneered and explained in previous reports, is to form scatterplots of observed vertical wavelengths versus theoretical hydrostatic vertical wavelengths given by the local background wind speeds, i.e. $\lambda_z = 2\pi|c-U|/N$, where c is the ground-based phase speed, U is the background wind along the wave vector direction, and N is the Brunt-Väisälä frequency. Figure 36a plots such a scatterplot of values from the equatorial region in and around the south China sea region, with Figure 36b showing the mean amplitudes of each point

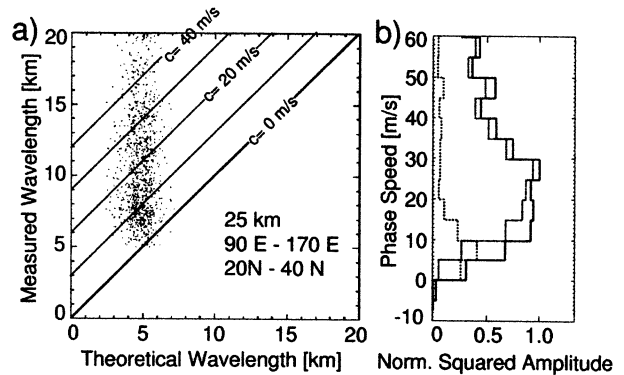


Figure 36: (a) scatterplot of theoretical versus measured vertical wavelengths and (b) amplitude of various points in (a) as a function of their inferred phase speed c . For more information, see Preusse *et al.* [2000d].

binned as a function of its inferred phase speed c from Figure 36a using the formula above (isolines of these inferred c values are also plotted in Figure 36a). We see inferred wave phase speeds peaking at ~ 20 - 30 m s^{-1} , although the peak is not sharp and larger values are also evident. Figure 37 shows a detailed profiling of some of this activity in the CRISTA2 data over regions

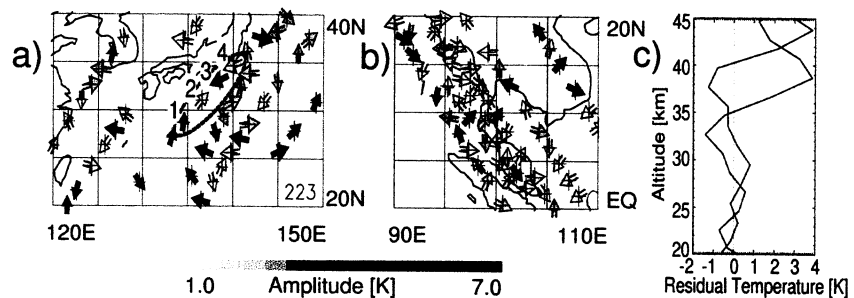


Figure 37: Gravity wave amplitudes and phases (a) east of Japan and (b) north of Indonesia during CRISTA2. Two adjacent profiles from panel (b) are plotted in panel (c), showing the long wavelength gravity wave and a phase shift $\sim 180^\circ$ between profiles, consistent with a horizontal wavelength ~ 300 - 400 km.

near Japan and Indonesia, using a plot formulation similar to that in Figure 27b. We see lines of activity along orbital tracks that show amplitude and phase coherence over several measurements, such as the sequence labeled 1-4 in Figure 37a. The wavelengths and phase shifts are also reproducible and consistent from profile to profile, as shown in an example in Figure 37c.

More complete versions of this work have now been published [Eckermann *et al.*, 2000b; Preusse *et al.*, 2000d] and the interested reader should refer to both of those papers for more information. In summary, however, this work has revealed stratospheric gravity waves in the equatorial and subtropical regions of the stratosphere that appear to correlate with increased water vapor abundances in the high tropical troposphere, which we interpret here as a proxy for increased convective activity. Thus, we associate these waves with tropical convective sources. Whether the dominant source is isolated clusters of convection in the region (Figure 35c), or whether the large coherent typhoon system in Figure 34 is the dominant source, is not yet clear and remains an area of ongoing research.

2.3.5 Gravity Waves in the Mesosphere

Here we briefly summarize this subproject work, which has provided information on gravity waves in CRISTA temperatures from high altitude scan (HAS) data. Here temperatures are derived from the $15\mu\text{m}$ CO_2 channel [e.g., Riese *et al.*, 1999]. Some of this work has already been presented, both here (e.g., the upper level data in Figures 10c and 10d come from the HAS data) and in previous reports (see, e.g., section 1.1. of the report dated September 18-December 18, 1998). Here we summarize one important new finding that has emerged from the analysis so far, namely what we believe to be the first space-based detection of gravity wave interactions with tides, which is one of the more important dynamical processes that occur at these altitudes.

Earlier work with the CRISTA mesospheric temperature data by Ward *et al.* [1999] and Oberheide *et al.* [2000] revealed a large well-defined diurnal tidal signature in these mesospheric temperatures during CRISTA1. It is well-known from ground-based measurements and models that tides can significantly modulate gravity waves, though this has never been observed from space. To investigate this here, we separated the CRISTA1 data into ascending and descending nodes, and performed separate Kalman fits to each. Since the local times of these data are ~ 12 hours apart, this fit goes some way to also fitting the diurnal tidal structure globally. The top row of Figure 38 shows latitude-height results of our MEM/HA analysis of perturbations in these separated ascending (daytime: $\sim 9\text{am}$ LT) and descending (nighttime: $\sim 9\text{pm}$) residual temperature fields, in the vertical wavelength range of $\sim 6\text{--}9$ km, which we can be fairly sure is not contaminated by any remaining tidal perturbations missed by the Kalman detrending.

There are large differences between the daytime and nighttime data, particularly in the equatorial regions where the diurnal tidal temperature oscillations are large [Ward *et al.*, 1999]. From the Hough mode structures here, we expect the horizontal velocity oscillations to be moderate. Thus, it seemed possible that the temperature perturbations of the tide might be the dominating influence at the equator, modulating the background temperatures and Brunt-Väisälä frequencies seen by the gravity waves and potentially producing these differences in observed gravity wave activity in the CRISTA1 HAS data.

To investigate this, we note from spectral gravity wave models and theories that waves in the wavelength range $6\text{--}9$ km should lie in the nominal “saturation” regime, since they have scales less than the characteristic wavelength λ_z^* , which is $\sim 15\text{--}30$ km in the mesosphere [e.g., Smith *et al.*, 1987]. Saturated temperature variance in this range should then satisfy the approximate relation $T'/(\bar{T}N^2) = \text{constant}$, where the constant is determined by the intensity of the saturated m^{-3} portion of the spectrum here [Smith *et al.*, 1987]. Thus, tidal modulations of both N^2 and \bar{T} profiles presented to gravity waves may cause T' amplitudes to vary so as to keep the saturation ratio $T'/(\bar{T}N^2)$ constant. To test this idea, we plot $T'/(\bar{T}N^2)$ in the bottom row of Figure 38 using the Kalman fitted background profiles in each

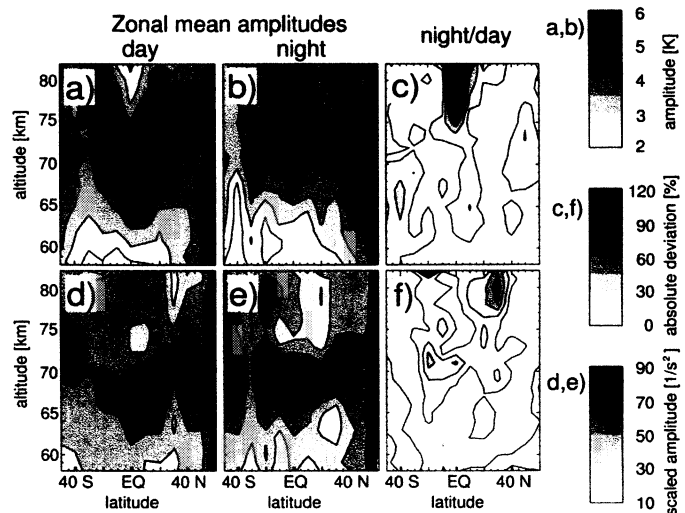


Figure 38: Top row: comparisons of daytime and nighttime zonal mean gravity wave \hat{T} values from the $15\mu\text{m}$ upper stratospheric and mesospheric temperature profiles obtained by CRISTA during November, 1994. Bottom row shows plots after rescaling of results by $\bar{T}N^2$, with values expressed now as percentages throughout. The day-night differences in (c) largely disappear after this normalization.

case. Indeed, we see that this ratio, while not a perfect constant, is much less variable than the temperatures in the top row of Figure 38.

Thus, we believe that Figure 38 provides initial evidence of gravity wave-tidal interactions in the equatorial mesosphere produced by strong diurnal temperature variability due to the diurnal tide. This work, along with additional analysis and discussion, has recently been published by *Preusse et al.* [2000b].

2.4 Gravity Wave Activity in MLS Radiance Data from UARS

Gravity waves have previously been observed in the microwave limb radiances measured by the Microwave Limb Sounder on UARS [e.g., *Wu and Waters*, 1997]. Initial controversy centered on why these zonal-mean MLS variances differed so drastically from similar temperature-related variances derived from other satellite instruments, such as LIMS and GPS/MET. Modeling work by *Alexander* [1998] suggested that the spatial sensitivities of each satellite instrument were the key determinant in setting spatial patterns in their zonal-mean variance maps. In section 2.2 we outlined the results of a detailed study of ours, using our “gravity wave calibrated” subset of CRISTA1 data as a basis, which proved experimentally the modeling hypothesis set forth by *Alexander* [1998]. In particular, we were able to use the vertical sensitivities of both CRISTA and MLS to convert CRISTA temperature data into a zonal-mean distribution that should be more like that which MLS would resolve. When compared to actual MLS variance maps in Figure 10, the comparisons were much improved. The conclusion is that MLS did indeed measure gravity waves reliably during the UARS mission period, but that it measured much longer vertical wavelength gravity waves than did other limb sounders, such as CRISTA, LIMS and GPS/MET.

Another consequence of this finding, reinforced in a subsequent climatological analysis of MLS data at 38 km by *McLandress et al.* [2000], was that longitudinal variations in MLS radiance variances at a given latitude should not be greatly influenced by instrument observing effects, and thus any zonal asymmetries in MLS radiance variance should be a reliable indicator of gravity wave variability of some sort or another. Figure 39 plots some mean results of MLS variances during winter in the southern and northern middle latitudes, as inferred from limb-track radiances [after *McLandress et al.*, 2000]. These limb track radiances essentially “stared” at a given atmospheric location, yielding spatially distributed data along the orbital track. The radiance measurement in each channel saturated at a certain different altitude, somewhat above the tangent point. These limb track data differ from the conventional limb-scanned data that were acquired exclusively by MLS earlier in the UARS mission. The idea behind the limb track mode was to “save time” by not scanning vertically and just tracking a given limb location. Thus these saturated limb-track radiances could be acquired much more rapidly along track (MLS viewed at a 90° angle to the spacecraft motion), yielding much greater horizontal resolution in the data, and potentially resolving a greater portion of the gravity wave spectrum than before in the limb scan mode. For further information, see *McLandress et al.* [2000] and *Jiang and Wu* [2000].

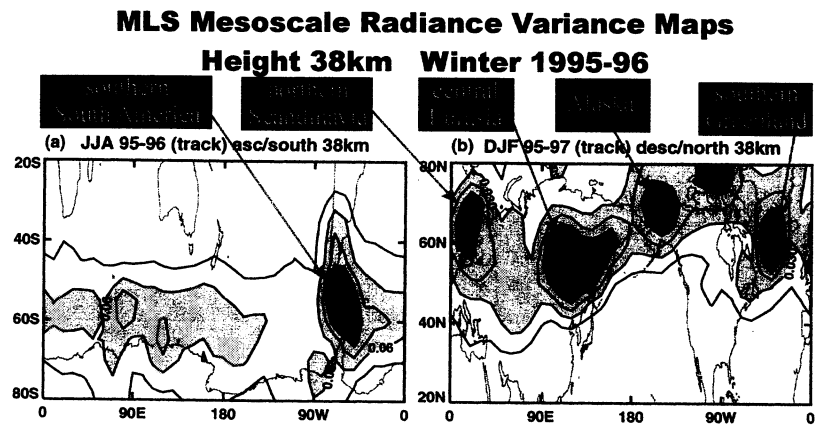


Figure 39: Maps of MLS limb-tracking radiances, averaged during winter months of 1995-1996 [after *McLandress et al.*, 2000]. Geographical locations of evident variance enhancements are pointed out.

In Figure 39a we see a large climatological peak in limb track radiances over the southern Andes, which resembles the one seen in CRISTA data (see Figures 14,15,17,20,21,27 and 30). *McLandress et al.* [2000] used a simple analytical model to argue that the peak in Figure 39a was probably due to gravity waves with ground-based phase speeds c clustered near zero, which suggests a mountain-related source.

In collaboration with Drs. Jiang and Wu from JPL, we started looking in depth at daily time series of MLS radiance variances and radiosonde fluctuations from this region of the southern Andes. Figure 40a shows monthly mean values of MLS radiance variance over the southern Andes from limb-tracking data acquired over the period 1995-96. The plot below (Figure 40b) shows results from 6 radiosonde sounding stations scattered around southern South America, from which temperature fluctuations were extracted above 100 hPa using standard analysis techniques [e.g., *Allen and Vincent*, 1995]. We see that there is a clear annual variation in the stratospheric wave activity over the southern Andes, with a large peak in winter and a deep minimum during summer months.

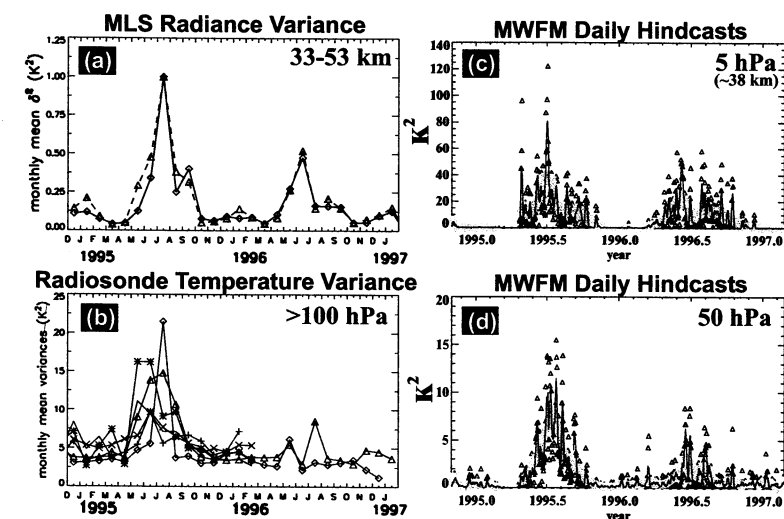


Figure 40: Mesoscale temperature variance in the stratosphere over the southern Andes region from November 1994 to February 1997. Left column shows data: (a) monthly mean MLS radiance variances over the Andes region (60°-80°W, 30°-50°S) at heights ~33-53 km. Note that MLS radiance variances are proportional to, but significantly smaller than, actual temperature variance. Separate curves show data with slightly different horizontal wavelength filter cutoffs; (b) Temperature variances above 100 hPa from six radiosonde stations in the southern Andes region. Right column shows MWFM hindcasts over the southern Andes region (50°-80°W, 40°-46°S) using daily (12Z) DAO assimilated data at (c) 5 hPa and (d) 50hPa. These altitudes roughly coincide to those for the data plots on the left. Triangles show daily means of the MWFM hindcasts, dark curve shows 5 point running average, thick light curve shows 30 point running average.

detailed MWFM hindcast simulations appear to prove beyond reasonable doubt that, not only is the climatological peak over the Andes in Figure 39a due to stratospheric mountain waves, but also most of the day-to-day variation in MLS radiances over this region year-round can be explained in terms of mountain waves transmitted into the stratosphere by flow over the southern Andes.

We also note in Figure 39b that some of the other peaks in the Northern Hemisphere, which *McLandress et al.* [2000] had difficulty in associating with a clear source and/or instrument effect, appear to correlate with major mountain ranges. For example, long wavelength stratospheric mountain waves over northern Scandinavia have been observed and modeled in a number of previous studies [e.g., *Dörnbach et al.*, 1998]. CRISTA1 data showed enhanced stratospheric temperature amplitudes over

To see if these results could be explained solely in terms of seasonal variations of stratospheric mountain wave activity from the Andes, we conducted an intensive set of MWFM hindcasts over this southern Andes region using DAO wind and temperature analyses for every day from October 1994 to March 1997, a total of ~800 individual hindcast runs in all. The results were binned and averaged as a function of height. Final daily average results over the same latitude and longitude range are plotted at 5 hPa (Figure 40c) and 50hPa (Figure 40d). These hindcast simulations show remarkable similarities to the MLS and radiosonde data. The seasonal variation is accurately reproduced. Furthermore, the stronger peak in 1995 compared to 1996 is also reproduced. Some of the month-to-month variability also appears to be captured in the simulations, such as the minor “notch” in activity during June-July, 1996. These highly

central Eurasia and Alaska, both of which were modeled in an MWFM hindcast simulations (Figures 18 and 19 – see also Figure 4 of *Eckermann and Preusse* [1999]). Intense stratospheric mountain wave fluctuations over southern Greenland were also measured by the MMS on an ER-2 flight on January 6, 1992, a case forecast by MWFM [*Bacmeister et al.*, 1994] and recently simulated in detail using a three-dimensional mesoscale model by *Leutbecher and Volkert* [2000]. Thus, much of the climatological variability in middle-to-high latitude MLS radiance variance in Figure 39b may be due to mountain waves. This idea will be fully assessed and tested in future multiyear MWFM hindcast simulations along the lines set forth in Figure 40c,d. The current MLS work for the Andes is currently being prepared for publication [*Jiang et al.*, 2000].

2.5 Gravity Wave Activity in CLAES Temperatures Measured from UARS

2.5.1 Motivation

To date, gravity waves have been detected in data from only a few instruments on the UARS spacecraft. Of these, data from MLS are the best known and most extensively analyzed to date [*Wu and Waters*, 1997; *Alexander*, 1998; *McLandress et al.*, 2000]. In the mesosphere and lower thermosphere gravity wave signals have been detected in various measurement channels of the HRDI and WINDII instruments [e.g., *Wang et al.*, 2000]. However, stratospheric gravity waves do not appear to have been seen to date from other major stratospheric probing instrument on UARS.

We conducted a detailed pilot to study to apply what we have learned from analysis of CRISTA temperatures for gravity waves, summarized extensively in sections 2.1-2.3 of this report, to see whether there is gravity wave content in stratospheric temperatures measured by the Cryogenic Limb Array Etalon Spectrometer (CLAES) on UARS. We have done this because CLAES is an instrument similar in concept and operation to CRISTA: like CRISTA, CLAES is a cryogenically cooled limb-scanning infrared spectrometer, acquiring spectra within a more limited spectral range of 3.5-12.9 μm . Unlike CRISTA, which acquired full infrared spectra, CLAES used a series of filters on a chopper wheel to isolate lines within specific narrow regions of interest [*Roche et al.*, 1993]. Nonetheless, like CRISTA, CLAES temperatures in the stratosphere and lower thermosphere were inferred from the CO_2 Q-branch limb emissions at $\sim 792\text{ cm}^{-1}$ (12.6 μm).

Gille et al. [1996] provided an extensive validation study of the CLAES temperature product. Comparisons with global analyses and radiosondes showed quite good agreement, with differences of $\sim 2\text{ K}$. *Gille et al.* [1996] estimated random statistical errors in Version 7 CLAES temperatures to range from $\sim 0.9\text{ K}$ at 100 hPa ($\sim 16\text{ km}$) to $\sim 2.2\text{ K}$ at 0.46 hPa ($\sim 53\text{ km}$). The vertical resolution of the profiles is $\sim 2.5\text{ km}$, giving a Nyquist vertical wavelength of $\sim 5\text{--}6\text{ km}$. While CRISTA temperatures are a little more accurate than this (estimated typically in the 0.5-1 K range), our analysis revealed stratospheric

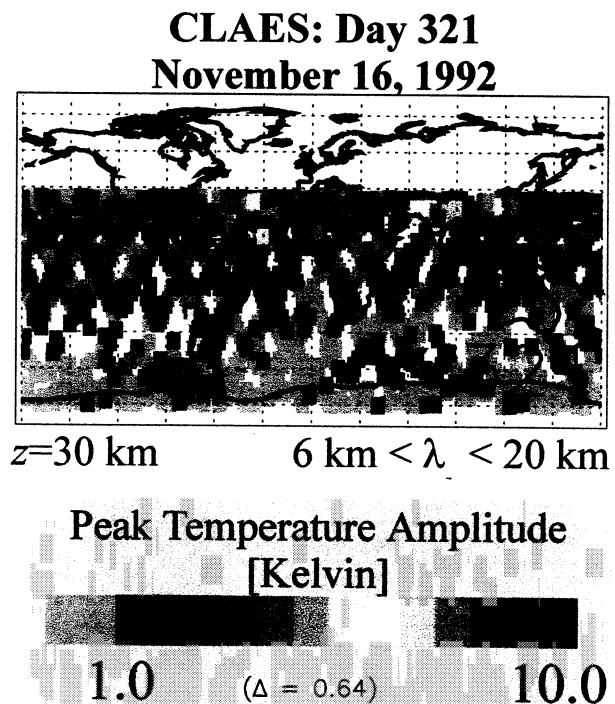


Figure 41: Global maps of gravity wave MEM/HA peak temperature amplitudes from CLAES measurements at 30 km altitude for 2 days starting on Day 321 of 1992 (November 16-17), with results plotted for fluctuations in the vertical wavelength range between 6 and 20 km.

gravity wave temperature perturbations as large as 10 K in the CRISTA data (Figure 14). Thus, on the basis of our experience with the CRISTA temperature data and the estimated random errors in CLAES temperatures, there seemed good reason to believe that stratospheric gravity waves may be evident in CLAES temperature data, since the along-limb acquisition characteristics should be broadly comparable. Indeed, stratospheric Kelvin waves have been seen in equatorial CLAES temperatures [Shiotani *et al.*, 1997; Canziani, 1999].

2.5.2 Subset of Southward-Viewing CLAES Data: November, 1992

Accordingly, we acquired a subset of the latest version (Version 8) CLAES level 3AT data from the Goddard DAAC. We concentrated first on data from November-December 1992, since this period corresponds best to the November, 1994 period of the first CRISTA-SPAS mission. Since we have a good understanding of the global morphology of gravity wave temperature variance from this mission, it provided a good initial basis from which to analyze and interpret similar potential signals in CLAES temperatures.

In this spirit, we analyzed the CLAES temperature fields in exactly the same way that we analyzed the CRISTA data, fitting the full day's data using a wavenumber 0-6 Kalman filter, subtracting to leave a small-scale temperature residual, then analyzing residual temperature fluctuations using a Maximum Entropy (MEM) to locate wavenumber peaks, and harmonic analysis (HA) to delineate vertical variations in peak wave amplitudes, as set forth for CRISTA in section 2.3.1. MEM/HA estimates at each altitude are stored and global maps for each day and plotted accordingly.

Figure 41 shows temperature perturbation amplitudes \hat{T} at 30 km altitude derived from the primary MEM/HA spectral peak in the CLAES temperature profiles acquired from UARS over a two day sequence starting November 16, 1992. Figure 14 shows a similar pattern of wave amplitude estimates from CRISTA on 6 November, 1994.

We note on comparing Figures 14 and 41 that the geographical coverage of each instrument is rather different in each case. This is caused by differences in the viewing geometry of each instrument on the spacecraft. Unlike CRISTA, which was backwards viewing, CLAES viewed the limb at a 90° angle to UARS' orbital motion. Thus, CLAES viewed the atmosphere farther to the north and farther to the south than CRISTA (to latitudes ~80°) in one hemisphere, depending on the yaw cycle, but has much more limited coverage of the opposite hemisphere (only to latitudes ~34°). The sampling characteristics are very much like those of the MLS instrument [see, e.g., *McLandress et al.*, 2000]. We see from

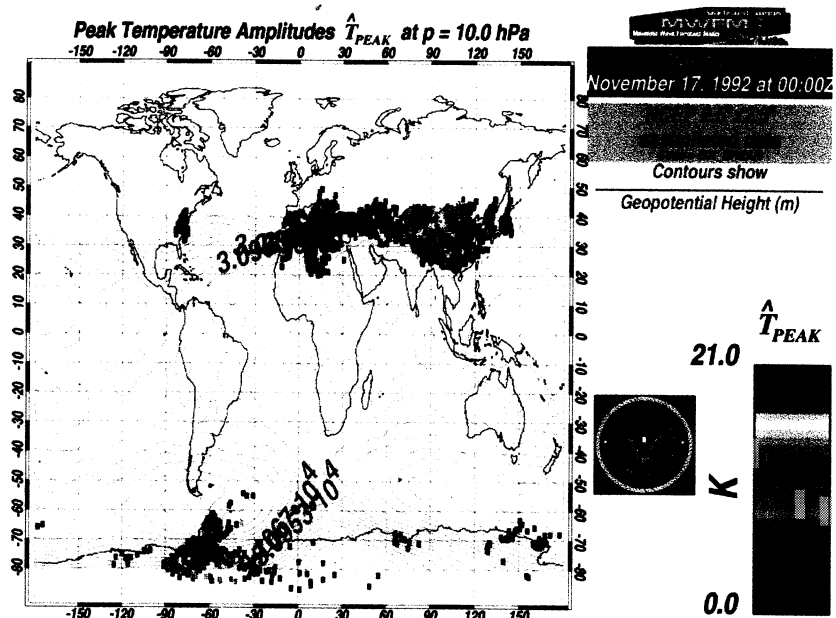


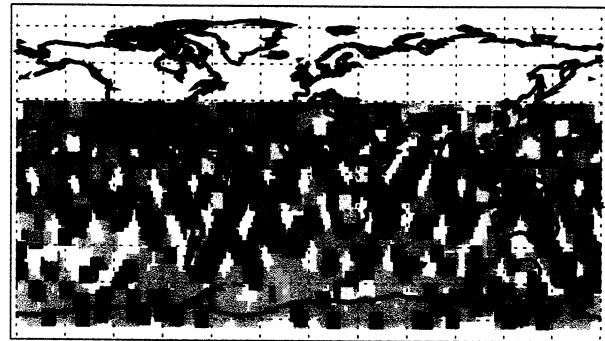
Figure 44: MWFM hindcast within the CLAES view area for southward UARS yaw cycle (~35°N-80°S) on November 17, 1992 at 0UT, at an altitude of 10 hPa (~32 km). Pink contours show geopotential height contours at this pressure altitude. This hindcast was constrained by the NCEP 2.5°x2.5° reanalysis fields for this time and date. Results were filtered to remove horizontal wavelengths > 100 km and vertical wavelengths > 5 km, to mimic the anticipated selectivity of the CLAES temperature data to long wavelength gravity waves.

Figure 41 that, on November 16-17, 1992, the UARS yaw cycle was such that CLAES preferentially viewed the Southern Hemisphere.

With these sampling differences in mind, we see qualitative similarities between the CLAES November 16-17, 1992 observations in Figure 41 and the CRISTA observations on November 6, 1994 in Figure 14. First, large regions of the stratosphere over the Southern Ocean have very small \hat{T} values. In equatorial regions, however, typical ambient \hat{T} values are somewhat larger.

More apparent, however, is the large isolated region of enhanced activity located directly above the southern Andes. Extensive analysis outlined in previous reports and publications [Eckermann and

CLAES: Day 321
November 16, 1992



$z=40$ km $6 \text{ km} < \lambda_z < 20 \text{ km}$

Peak Temperature Amplitude
[Kelvin]

1.0 ($\Delta = 0.64$) **10.0**

Figure 42: As for Figure 41, but at an altitude of 40 km.

CLAES: Day 321
November 16, 1992



$z=50$ km $6 \text{ km} < \lambda_z < 20 \text{ km}$

Peak Temperature Amplitude
[Kelvin]

1.0 ($\Delta = 0.64$) **10.0**

Figure 43: As for Figure 41, but at an altitude of 50 km.

Preusse, 1999; see sections 7.1 and 7.3] has definitively associated the CRISTA mesoscale temperature enhancement over the southern Andes in Figure 14 as being due to a long mountain wave forced by flow across the Andes (see, e.g., Figures 15,17, 21,27).

Also apparent in Figure 41, but less apparent in Figure 14, is a large cluster of large \hat{T} values to the north of India. This is interesting since it corresponds to the high mountainous regions of the Himalayas. These regions of enhanced temperature amplitudes stand out from other regions at similar northern latitudes $\sim 30^\circ\text{N}$, where much reduced \hat{T} values are evident. All of this is strong (but preliminary) evidence that there are the manifestations of long wavelength gravity waves radiated from these mountainous regions, which propagate into stratosphere and produce explicitly resolved temperature perturbations on CLAES temperature profiles.

To assess this, next we investigate the fluctuations in the CLAES data at higher altitudes in the stratosphere. If the fluctuations at 30 km in Figure 41 were due to a chance clustering of temperature errors in some profiles at 30 km, we might expect to see a somewhat different uncorrelated pattern 10 km higher at 40 km. Figure 42 shows the corresponding \hat{T} estimates at 40 km. Figure 43 shows the corresponding \hat{T} estimates at 50 km. The geographical distributions are very similar to those seen in Figure 41, which is inconsistent at least with random errors in CLAES temperatures that are uncorrelated in altitude, and consistent with a geophysical interpretation in terms of gravity wave amplitude trends that persist as waves propagate from 30 km to 40 km and then to 50 km in altitude.

To test whether these might be mountain waves, we conducted MWFM hindcasts for this period, based on recently issued $2.5^\circ \times 2.5^\circ$ NCEP global fields from the 40 year reanalysis project [Kalnay *et al.*, 1996]. We also conducted comparison simulations with DAO data for the period, which consisted solely of the coarser $5^\circ \times 4^\circ$ analysis fields. The results were basically comparable, though they showed some differences that we ascribe to shortcomings in the coarser and older DAO assimilations, possibly in the troposphere where early DAO analyses were believed to have problems. For present purposes, we use the NCEP analyses as our “best guess” for what the atmosphere was doing at this time.

Figure 44 shows results from MWFM hindcast runs constrained by the NCEP reanalysis fields for 17 November 1992 at 0 UT, a time roughly at the center of the 2 day CLAES observation period November 16-17 (Figure 41). These hindcast simulations predict a concentration of enhanced mountain wave temperature variance in the stratosphere over the extended Himalayan Region, much as observed in Figure 41 and at higher altitudes in Figures 42 and 43. Further, other mountainous regions at a similar latitude, such as the Rocky Mountain Region over the western USA, produces no detectable gravity wave variance, again much as observed in Figures 41-43. The absence of gravity waves here in both model predictions and data is, in some ways, even more persuasive, since it agrees with what we know about mountain wave propagation: i.e., on November 16, 1992, MWFM/NCEP hindcasts suggest that mountain waves could freely propagate into the stratosphere with long wavelengths over the Himalayas, where they could be detected by CLAES. Conversely, mountain waves were either not strongly forced, were blocked from propagating into the stratosphere, or had wavelengths too small to be resolved by CLAES, and so should not have been detected. Thus, detection over the Himalayas and non-detection over the western USA, as simulated and observed, is, in a sense, double confirmation that this activity over the Himalayas in CLAES temperatures is a geophysically believable gravity wave signature.

There are, however, some discrepancies also. For example, the isolated peak over the southern Andes, evident in

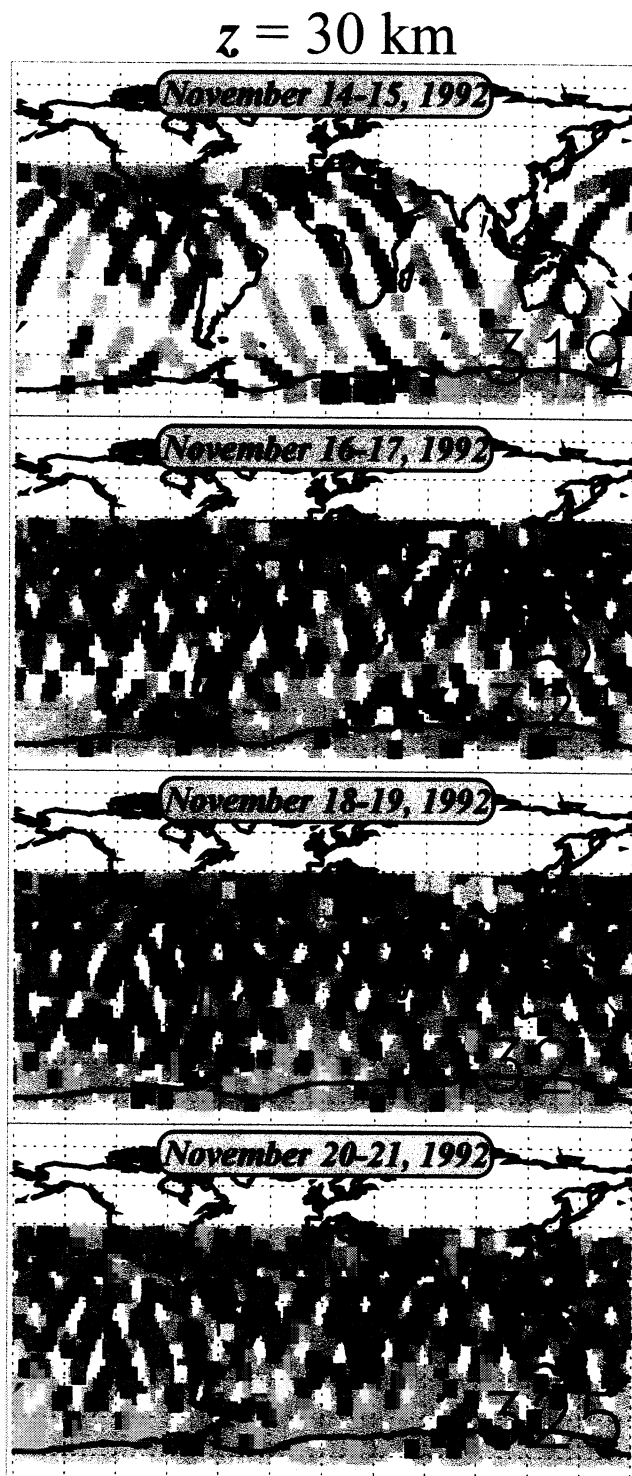


Figure 45: As for Figure 41, but showing a sequence of 2-day maps of CLAES \hat{T} amplitudes at 30 km, from November 14-21, 1992 at an altitude of 30 km.

Figures 41-43, does not show up in the MWFM hindcast in Figure 44. Hindcasts for other times and days, and in which wavelength filtering was not applied to the hindcast output, did reveal an occasional small isolated peak over the southern Andes, but usually of fairly small amplitude (not shown). To study this a bit more, Figure 45 shows results a sequence of CLAES \hat{T} plots at 30 km for the day pairs immediately before and after those shown in Figure 41 (which is reproduced in the second panel from the top in Figure 45). The sampling on the preceding days (top panel) is a little sparse globally, so it is hard to discern trends on the days prior. For the days afterwards (bottom two panels in Figure 45), however, we note that we have better global coverage in the CLAES data. These plots show that the extended region of activity over the Himalayan Region persists and evolves somewhat over the following four days (18-21 November, 1992), giving us further confidence in the fidelity and geophysical reality of this patch of variance in the CLAES temperature data.

The “peak” over South America, however, is much more fickle and does not seem to persist as strongly with time in Figure 45. Assuming it is real (i.e. geophysical), there are several plausible reasons for explaining how such waves might arise in the CLAES data but not be simulated by the MWFM hindcasts. One possibility is weaknesses in the assimilated datasets that are being used here. Since these fields are reanalyses based on assimilated data acquired about a decade earlier, they may not accurately simulate fine scale flow features in and around this region of the far southern Andes that may be important for the forcing and propagation of long mountain waves into the stratosphere at the time. Another plausible reason may be that the MWFM algorithms may have weaknesses here. Many processes may be important for forcing mountain waves in this region, but are ignored currently by the simplified analytical parameterization algorithms at the core of the MWFM. One potentially important effect is time dependence of the background wind conditions. MWFM (as with most mountain wave schemes) assumes a time-independent background atmosphere, whereas, in fact, wind conditions tend to vary considerably as a function of time as weather systems move across the topography and evolve. These and other possibilities are a prime focus of planned future investigations of these data.

2.5.3 Subset of Northward-Viewing CLAES Data: December, 1992

Given the promise shown in section 2.5.2 analyzing the southward-viewing subset of CLAES data from November, 1992, we extended the analysis to some northward viewing CLAES data acquired during December, 1992, by which time UARS had maneuvered to its opposite yaw phase, and thus CLAES was viewing the atmosphere from the other side of the spacecraft’s orbital track. This yields measurements which preferentially probe the middle atmosphere in the latitude band from $\sim 35^\circ\text{S}$ to $\sim 80^\circ\text{N}$, giving a northern hemisphere bias to the data. A global map of CLAES peak residual temperature amplitudes \hat{T} on December 20-21, 1992, inferred from the MEM/HA analysis at an altitude of 30 km are plotted in Figure 46, with results shown for fluctuations with vertical wavelengths in the range ~ 6 -20 km.

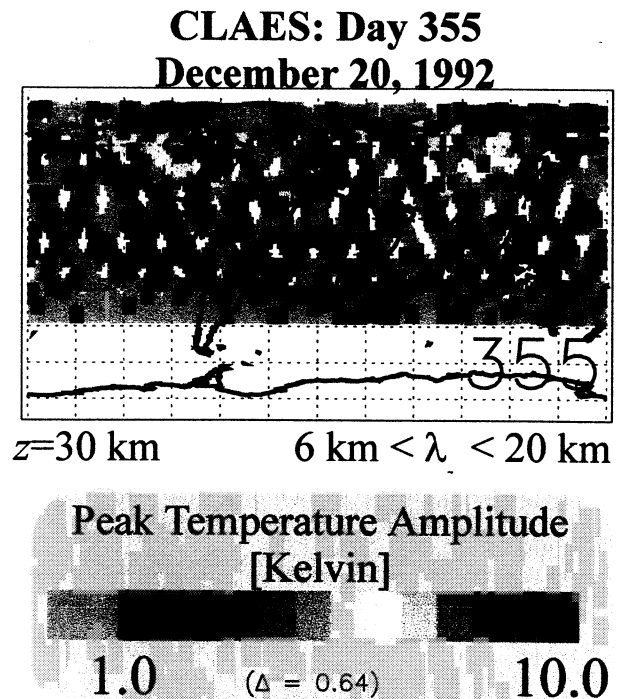


Figure 46: As for Figure 41, but for CLAES data at 30 km starting on Day 355 of 1992 (20-21 December) during the opposite UARS yaw cycle, which led CLAES to preferentially measure the northern hemisphere rather than the south, as evident from the figure.

As for the November data in Figure 41, we see a preferential clustering of the largest temperature amplitudes over the major land masses. In this case, notable regions are the high-latitude regions of North America, southern Greenland, northern Scandinavia, north of India and eastern Eurasia. As is well known, of course, all of these regions are not just major land masses, but have significant mountain ranges situated there. We note similar features in the climatological MLS limb-track radiance variances at 38 km in Figure 39b.

Figure 47 shows an MWFM hindcast simulation at 10 hPa, based on NCEP reanalysis winds and temperatures for December 20, 1992 at 12 UT. Results are filtered to retain only the long wavelength mountain waves that the CLAES limb-scan data should theoretically be sensitive to, based on what we learned from CRISTA limb scans (see Figure 3). The MWFM hindcast in Figure 47 shows remarkable visual similarities to the CLAES peak temperature residuals plotted in Figure 46. In particular, the bands of enhanced variance over far northwestern and far northeastern North America, the southern tip of Greenland, northern Scandinavia and central Eurasia all show out in the MWFM hindcast. Perhaps more impressively, other regions where large mountain ranges exist which could also force mountain waves produce no discernable signal in either the CLAES data in Figure 46 or the MWFM hindcast in Figure 47 – examples include mid-western USA, and the tropical Andes regions along the west coast of South America.

Indeed, previous observational and modeling work has shown the episodic presence of long wavelength mountain waves in several of these middle and high latitude northern stratospheric regions noted above. For example, ground-based and airborne stratospheric observations in northern Scandinavia have frequently revealed long wavelength mountain waves in the lower stratosphere during winter which produce temperature variability that modulates and accentuates the formation of polar stratospheric clouds [Wirth *et al.*, 1999]. As mesoscale model simulations reveal, the long wavelength response is produced by flow over the long wide Norwegian Mountain Range that runs along the coast of Norway [e.g., Leutbecher and Volkert, 1996; Dörnbrack *et al.*, 1998; Wirth *et al.*, 1999]. Despite their well-known presence, significant temperature amplitudes and long wavelengths, these waves have heretofore never been detected (explicitly at least) from space-based stratospheric temperature profiling. They could not be seen in CRISTA data since CRISTA could not observe far enough north. Accumulated limb-track variances from MLS at 38 km in Figure 39b suggest some kind of signal in the broad vicinity, but no explicit modeling analysis of these features has been conducted to date. Thus, these apparent observations in Figure 46 may represent the first tentative observation of these long mountain waves from the Norwegian Mountains from UARS. Similarly, stratospheric mountain wave activity emanating from topography on the southernmost tip of Greenland were measured in stratospheric velocity fields at ~70 hPa by the MMS instrument on NASA's instrumented stratospheric

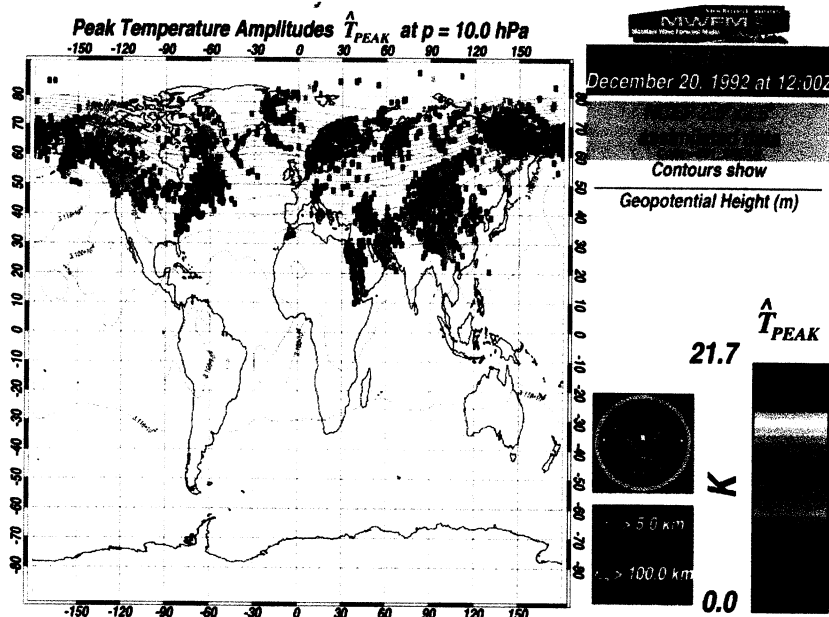


Figure 47: MWFM hindcast within the CLAES view area for northward UARS yaw cycle (~35°S-80°N) on December 20, 1992 at 12UT, at an altitude of 10 hPa (~32 km). Pink contours show geopotential height contours at this pressure altitude. This hindcast was constrained by the NCEP 2.5°x2.5° reanalysis fields for this time and date. Results were filtered to remove horizontal wavelengths > 100 km and vertical wavelengths > 5 km, to mimic the anticipated selectivity of the CLAES temperature data to long wavelength gravity waves.

ER-2 research aircraft on 6 January, 1992 [Chan *et al.*, 1993; see also <http://uap-www.nrl.navy.mil/dynamics/html/6jan92.html>]. Mountain wave forecasts [Bacmeister *et al.*, 1994] and hindcast simulations [Bacmeister *et al.*, 1994; Leutbecher and Volkert, 2000] definitively associated this activity with mountain waves and helped delineate the nature of the waves. In particular, the three-dimensional mesoscale model simulations of Leutbecher and Volkert [2000] revealed a great deal of structure in the simulated waves, showing both shorter and longer wavelength structures, the latter potentially resolvable from a satellite. Again, however, it does not appear that such waves have heretofore been detected from a satellite platform.

$$z = 40 \text{ km}$$

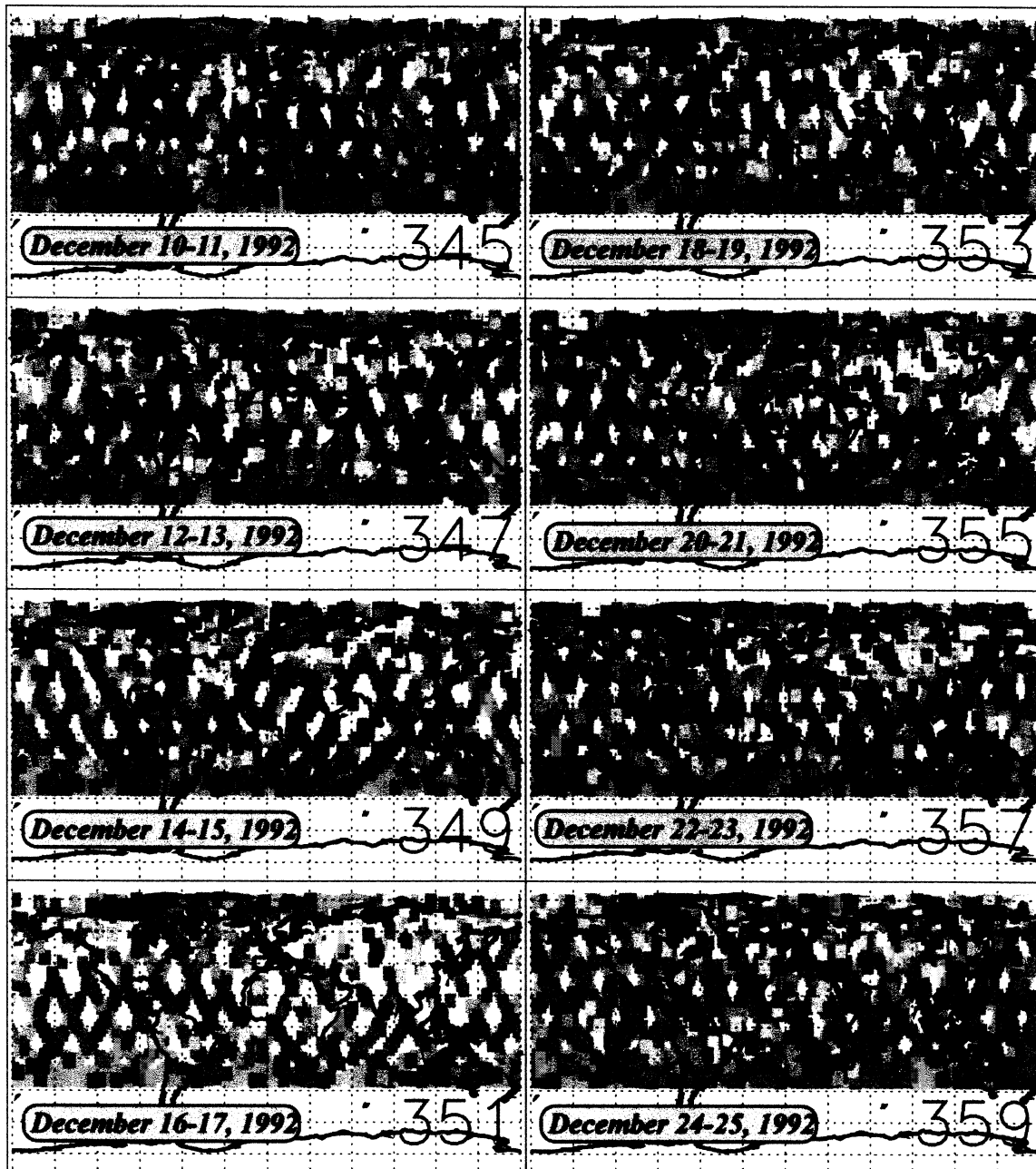


Figure 48: Sequence of global maps on successive day pairs of the peak amplitudes of CLAES temperature residuals, \hat{T} , at an altitude of 40 km.

Waves have been detected from space before over central Eurasia, however. Detections of such waves in CRISTA data in November, 1994, were reported in previous quarterly reports and published in the literature [Eckermann and Preusse, 1999; see sections 7.1 and 7.3]. These waves appeared to be radiated from complex topographic ridges along the border regions of China, Russia, Mongolia and Kazakhstan, although exact spatial correlations with associated MWFM hindcast simulations were not perfect. Again, the band of enhanced activity over central and eastern Eurasia in the CLAES \hat{T} plot in Figure 46 has many similarities to similar activity evident in the corresponding MWFM/NCEP hindcast map in Figure 47.

Figure 48 shows the global picture at a higher altitude of 40 km over a complete 16 day period of CLAES measurements in December, 1992 (Days 345-360). We see that activity over Northern America and Europe seems to be most intense during Days 353-358 (18-23 December), whereas in the eight days preceding (Days 345-353; December 10-17), the activity over North America largely disappears, while the activity over Europe, while it persists more, tends to be a little more dispersed. To assess this from the mountain wave perspective, Figure 49 shows sample MWFM/NCEP hindcasts at 10hPa for 12 December, 1992 (left panel) and 23 December, 1992 (right panel). Like the results in Figure 48, we see strong diminution of hindcast mountain wave amplitudes over North America earlier in December, with much of the continent free of forecast wave activity (Figure 49, left panel). Conversely, 10 days later (Figure 49, right panel), a great deal more of the North American has areas where stratospheric mountain waves are forecast. Hindcast differences over Europe are not as clear, however. The differences are also less distinct in the CLAES data in Figure 48, but there are indications of stronger activity later in December than earlier. This does not come out in the hindcasts in Figure 49, at least not obviously.

We believe that that this work has provided strong evidence of stratospheric mountain wave activity in the 12.6 μm temperatures measured by CLAES, which is an entirely new finding that was motivated by our analysis of a similar instrument (CRISTA) earlier in the project (sections 2.1-2.3). Thus, this adds another UARS instrument to the growing list that now appear to explicitly resolve gravity wave perturbations in one of their middle atmosphere channels. Our goal is to analyze these data further to provide information of a similar kind to CRISTA over the time periods and regions probed by CLAES before it exhausted its supply of onboard cryogen (\sim May, 1993). While this set of data is limited by UARS standards, it is much more extensive than the \sim 2 weeks of data provided by CRISTA, and the latter has kept us busy for many months of this contract. Thus, we believe there is much more to be done and found in this newly discovered CLAES gravity wave product.

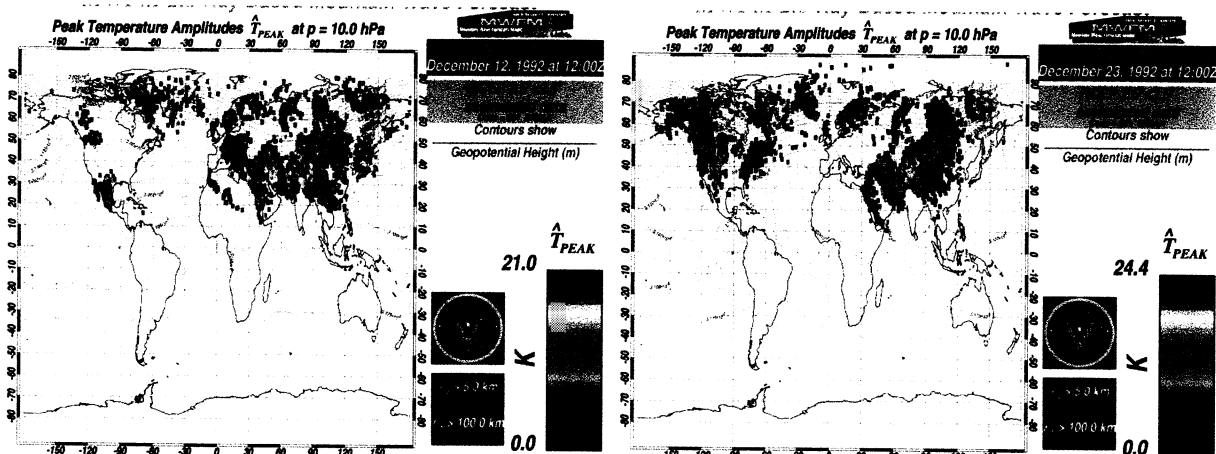


Figure 49: MWFM hindcasts constrained by NCEP reanalysis fields on December 12 (Day 347 - left panel) and December 23 (right panel: Day 358), at a height of 10 hPa and local time of 12 UT, with vertical wavelengths > 5 km and horizontal wavelengths > 100 km removed from the synthesized results.

3. Summary

3.1 Conclusions and Recommendations

Our work has proven that the era of routine space-based detection of gravity wave fluctuations in the stratosphere and mesosphere is at hand, a claim that was controversial and in doubt at the beginning of this project. Our work has provided insights into the precise way in which infrared limb scanning satellites explicitly detect long wavelength gravity waves in their profiled radiance data. These insights (e.g., section 2.1) will aid in the design of similarly motivated future instruments (e.g., HIRDLS) that can detect gravity waves by limb scanning, or via other observing strategies (e.g., nadir imaging).

We have conducted an extensive in-depth analysis of gravity waves resolved in the CRISTA limb-scanned temperature data (section 2.3), and have demonstrated that they provide extremely valuable global and precise geographical and vertical information on long wavelength gravity wave activity produced by a variety of sources, such as mountains, the jet stream, and tropical convection, as well as how these waves interact with tidal motions at higher altitudes. These insights have motivated further searches for such wave sources and characteristics in the MLS and CLAES temperature data acquired on UARS (sections 2.4 and 2.5).

Follow-up work is now needed to consolidate upon these findings. For example, we have shown that MLS and CLAES data resolve long wavelength gravity waves quite well, yet this rich vein of scientific information on stratospheric gravity waves has barely been tapped to date. Detailed analysis of these data, rather than the limited climatological analyses [McLandress *et al.*, 2000] or subset/subregion studies (sections 2.4 and 2.5) that have been attempted to date, is the logical next step forward. It is important that this is done, since our observational record on gravity waves in the middle atmosphere is still meager, yet these waves are known to be vital in setting the global circulation and climate.

3.2 Relation to Ultimate Objectives of the Research Contract

At the beginning of this contract, the very detection of middle atmospheric gravity waves explicitly from satellite platforms was still uncertain and controversial. Accumulated preliminary radiance/temperature variances from instruments such as LIMS and MLS clashed (see Figure 10), and, since these fluctuations were at the very detection limits (both in terms of accuracy/precision and spatial resolution), errors or contamination from larger scale structures (e.g., vortex filaments) could not be discounted. Detection of gravity waves from other instruments, such as infrared limb scanners such as CRISTA and CLAES, had yet to be studied and verified.

Thus, our goal in this research project was to cast light on space-based detection of gravity waves in the middle atmosphere, test these inferences using CRISTA data, and apply our knowledge to detecting and characterizing gravity waves that might be explicitly resolved in data from various other instruments on UARS. We made large strides in all of these goals during this project, as has been outlined in section 2.

Section 2.1 outlines the significant progress we made in understanding how infrared limb scanners measure gravity waves in the middle atmosphere. Section 2.2 illustrates why earlier zonal-mean data were somewhat discrepant and formulates methods for getting around this. Section 2.3 outlines a detailed analysis technique we developed to characterize long wavelength gravity waves in CRISTA data, which yielded new global information of stratospheric gravity waves generated by mountains (section 2.3.2), the vortex/jet-stream system (section 2.3.3), tropical convection (section 2.3.4) and the extension of these waves into the mesosphere where they interact with the diurnal tide (section 2.3.5). The detailed insights gained here were used to study climatological distributions of mountain waves evident in MLS data from UARS (section 2.4), and to prove for the first time that

stratospheric temperatures measured by the CLAES instrument on UARS also resolved stratospheric gravity waves (section 2.5). We believe that the full body of work summarized in this final report, and the ensuing publications listed in sections 4 and 5, have amply fulfilled all the research goals we set ourselves at the beginning of the project.

4. Papers Published in Peer-Reviewed Literature From Work Performed Under This Contract²

- Drob, D. P., J. M. Picone, S. D. Eckermann, C. Y. She, J. F. Kafkalidis, D. A. Ortland, R. J. Niciejewski, and T. L. Killeen, Mid-latitude temperatures at 87 km: results from multi-instrument Fourier analysis, *Geophys. Res. Lett.*, 27, 2109-2112, 2000.
- Eckermann, S. D. and P. Preusse, Global measurements of stratospheric mountain waves from space, *Science*, 286, 1534-1537, 1999.
- Eckermann, S. D., P. Preusse, and D. Offermann, Stratospheric gravity waves near the edge of the Southern Hemisphere polar vortex during the CRISTA2 mission, *J. Geophys. Res.*, (manuscript in preparation for CRISTA2 special issue), 2000a.
- Eidmann, G., D. Offermann, M. Jarisch, P. Preusse, S. D. Eckermann, and F. J. Schmidlin, Horizontal temperature variability in the stratosphere: global variations inferred from CRISTA data, *Adv. Space Res.*, (in press), 2000.
- Jiang, J., D. L. Wu and S. D. Eckermann, A study of mountain waves over the Andes: UARS MLS observations compared with MWFM numerical simulations, *J. Geophys. Res.*, (paper in preparation), 2000.
- Preusse, P., B. Schaeler, J. T. Bacmeister and D. Offermann, Evidence for gravity waves in CRISTA temperatures, *Adv. Space Res.*, 24, (11)1601-(11)1604, 1999.
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² Full listing of most of these papers, with hyperlinks to abstracts and electronic manuscripts, is available at the following web site: <http://uap-www.nrl.navy.mil/dynamics/html/puball.html>

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7. Appendix: Three Short Sample Publications Reporting Results from this Contract

7.1 Eckermann, S. D., and P. Preusse, *Science*, 286, 1534-1537, 1999 (4 pages)

7.2 Preusse, P., S. D. Eckermann and D. Offermann, *Geophys. Res. Lett.*, 27, 3877-3880, 2000. (4 pages)

7.3 Britt, R. R., *feature article in space.com*, November 18, 1999. (3 pages)

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